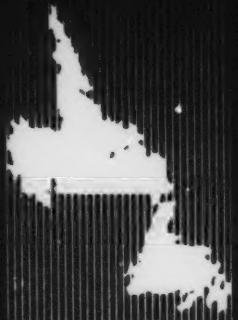

GEOLOGICAL SURVEY BRANCH
DEPARTMENT OF MINES AND ENERGY
GOVERNMENT OF NEWFOUNDLAND AND LABRADOR

CURRENT RESEARCH

REPORT 92-1

Compiled and Edited by
C.P.G. Pereira, D.G. Walsh and R.F. Blackwood

St. John's, Newfoundland
March 1992

CORRECTIONS

1. The photographs of Plates 4 and 9 of ''Ophiolites, sedimentary rocks, posttectonic intrusions and mineralization in the Eastern Pond (NTS 2D/11W) map area, central Newfoundland'', by W.L. Dickson, have been transposed; the captions are in the correct place.

2. The photographs of Plates 2 and 3 of ''Geological mapping in the Torngat Orogen, northernmost Labrador: preliminary results'', by R.J. Wardle *et al.*, have been transposed; the captions are in the correct place.

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EDITING, LAYOUT AND DRAFTING

Senior Geologist R.F. BLACKWOOD

Editor C.P. G. PEREIRA

Copy Editor D.G. WALSH

Typesetting B. STRICKLAND
J. MATTHEWS

Word Processing J. BANTON
D. DOWNEY
J. MATTHEWS
D. PEAVY

Supervisor, Cartographic AND Photomechanical Services K. BYRNE

Cartography G. DENIEF
D. LEONARD
T. PALTANAVAGE
T. SEARS

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PREFACE

This publication presents the preliminary results of the 1991 research program of the Geological Survey Branch of the Newfoundland Department of Mines and Energy.

Major features of the 1991 program include the initiation of a new bedrock-mapping project in northernmost Labrador, beginning of a pilot project to add geophysical data in mineral-assessment files to the digital geophysical database, start of a new Quaternary mapping project in western Labrador, release of a new microcomputer version of the Mineral Occurrence Data System (MODS), and publication of a new industrial-minerals map for the Island of Newfoundland.

About a third of the 1991 program was funded under the Canada–Newfoundland Cooperation Agreement on Mineral Development (1990-1994), which was signed in October, 1990. The Geological Survey of Canada also carried out a number of projects in the province under the Agreement. Full reports on these projects will be published in the GSC Current Research volume; we thank the GSC for permission to include the abstracts of these reports here.

B.A. Greene
Executive Director
Geological Survey Branch

Any reader who would like to write a rebuttal to, or discussion of, any report contained in this volume is invited to submit it to the editor by November 1, 1992, to be considered for inclusion in Report 93–1.

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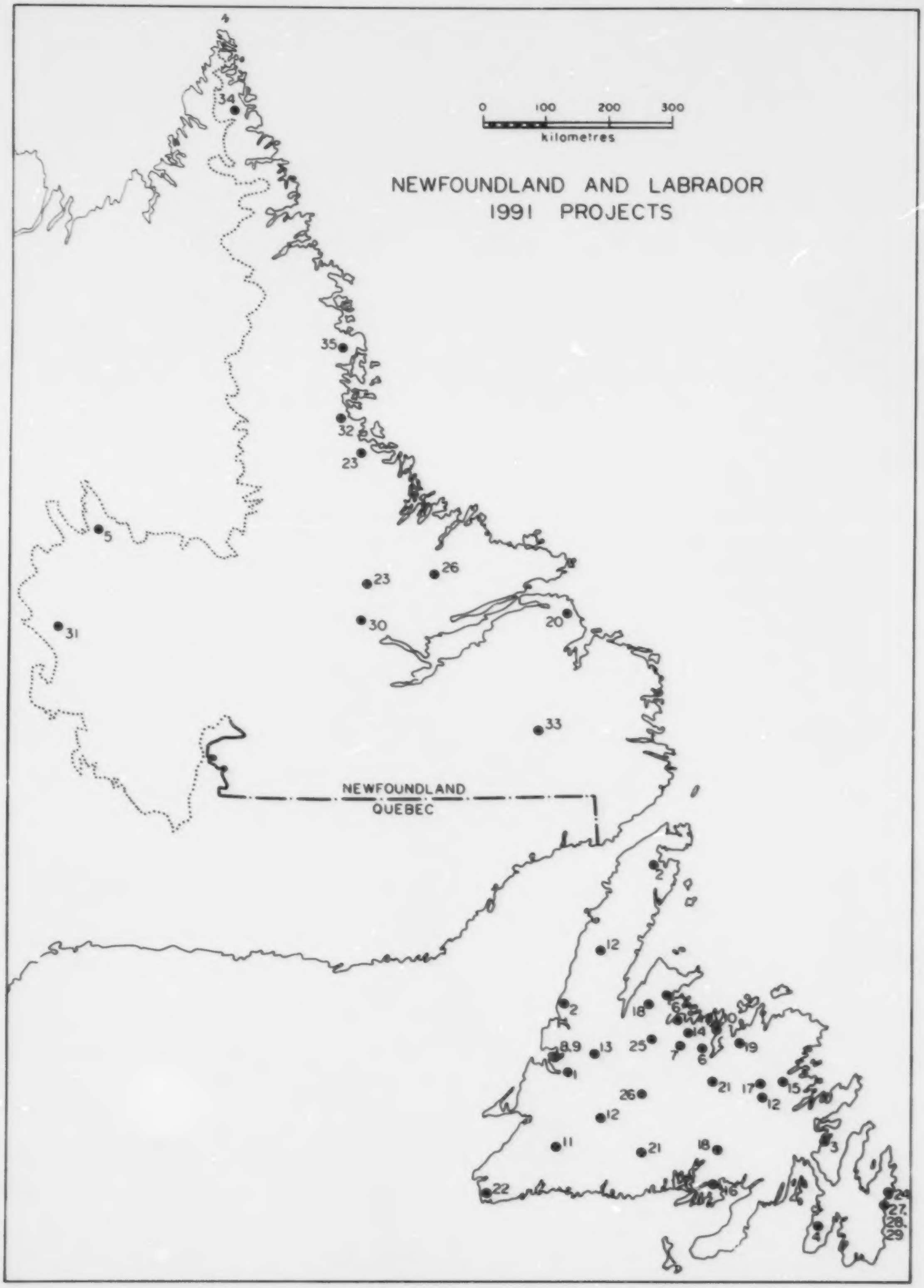

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*Current Research (1992) Newfoundland Department of Mines and Energy,
Geological Survey Branch, Report 92-1, pages 1-12*

QUATERNARY GEOLOGY OF THE CORNER BROOK–PASADENA AREA (NTS 12A/13 AND 12H/4)

M.J. Batterson and S.V. Vatcher
Terrain Sciences Section

ABSTRACT

Mapping of the Quaternary geology in the Corner Brook–Pasadena area defined both the regional ice-flow history and the distribution and characteristics of the surficial sediments. Such mapping is useful in a wide range of applications including engineering, town-planning, forestry and agriculture, as well as in mineral exploration.

Ice-flow data suggest an early southwestward flow down 'modern' Deer Lake toward the coast either through Wild Cove or the Humber gorge. As the ice mass developed, topography had less influence, and dispersal centres shifted to the Topsail Hills. This shift is indicated by the presence of clasts identified as originating from the Topsail Hills found in diamictons in the Deer Lake basin, and striation orientations.

Quaternary sediments are common across the area. On highlands underlain by metasedimentary bedrock, sediment is thin and discontinuous. Thicker sediment is common in valleys, and exposures containing more than one sediment unit are common. Generally, diamictons have similar characteristics, apart from colour, and are probably basal meltout tills. High-level deltas in the South Brook valley suggest the presence of a proglacial lake in the area. In the Humber valley, a typical sequence consists of a variable number of diamicton units, overlain by a succession of glaciofluvial, marine and postglacial fluvial sediments. Marine inundation occurred at least as far northeast as Cormack.

The wide variety of Quaternary sediments and their contrasting characteristics have implications for engineering (slope stability), land use (waste-disposal site location and their effects), hydrogeological (aquifer development) and resource management (agriculture, forestry) applications. Of the Quaternary deposits, marine muds are the most sensitive to disturbance.

INTRODUCTION

This report discusses results from the first year of a three-year study into the Quaternary geology of the Humber River basin. The glacial history of the area is complex because of the location of the basin between ice dispersal centres on the Long Range Mountains and on the Topsail Hills. As well, smaller late-glacial dispersal centres on Birchy Ridge and the southern Long Range Mountains potentially influence the region. During deglaciation, the area likely contained a large proglacial lake and, following lake drainage, was inundated by the sea at least as far northeast as Cormack. Exploration in areas of potential mineralization, such as for uranium in the Birchy Ridge area, has been hampered by the lack of adequate knowledge of the sequence of glacial events.

The range of sediment types, within the Humber River basin, resulting from the regions diverse Quaternary history have aided in the development of various resource-based activities, including agriculture and forestry. However, these same sediments also cause engineering and geotechnical difficulties, primarily in areas containing thick mud sequences.

The objectives of this project are to map the distribution of Quaternary sediments, to describe their characteristics and geomorphology, to map ice-flow indicators, and to develop a model for the glacial and postglacial history of the area. This will assist mineral-exploration activities by providing data on directions and possibly distances of transport of glacial sediment, and will also provide basic geological data of use to other projects influenced by surficial sediment.

LOCATION AND ACCESS

In 1991, the field area was most of the Corner Brook map sheet (NTS 12A/13), with the exception of Glover Island, and the southern half of the Pasadena map sheet (NTS 12H/4) (Figure 1). The field area contains the city of Corner Brook, and smaller communities including Massey Drive, Steady Brook, South Brook and Pasadena. Access was generally good either by using trucks on paved roads or ATV's on the numerous logging roads. Areas east of Corner Brook Lake and Glover Island were not examined, as they are only efficiently accessible by helicopter.

BEDROCK GEOLOGY

The use of bedrock geology to Quaternary studies is primarily in its role of aiding determination of distances and directions of glacial transport. Rock types that are visually distinctive and confined to a discrete source area are

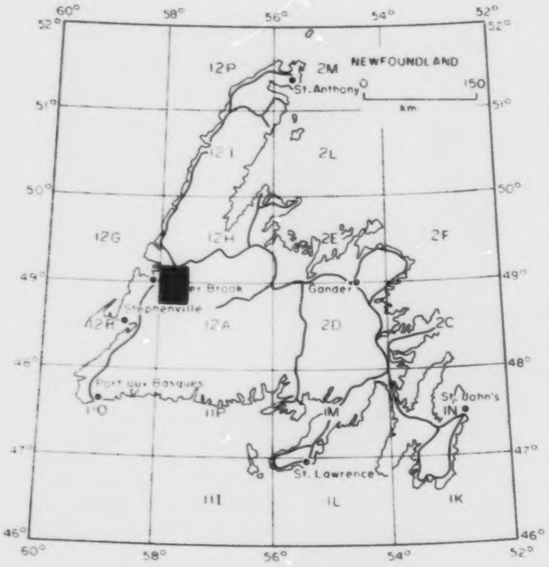

Figure 1. *Location of field area.*

particularly useful. The field area consists of Cambrian carbonates and sediments, Upper Proterozoic metasediments and metavolcanic rocks, and Carboniferous clastic sediments (Hyde, 1984; Williams and Cawood, 1989; Figure 2). Each of these rock types extend beyond the field area, and therefore, clasts from these units cannot be used for glacial-dispersal studies. The rock types surrounding the area are of greater potential use. To the north, and underlying that area affected by the Long Range ice cap, are Precambrian gneisses, whereas, the area underlying the Topsail Hills ice centre host granites and felsic volcanic rocks. Topsails granites are commonly peralkaline (Whalen and Currie, 1988).

PREVIOUS WORK

Reconnaissance surficial mapping of the Pasadena map sheet area was completed by Grant (1973) and of the Corner Brook map sheet area by Kirby *et al.* (1988). Batterson and Taylor (1990) described preliminary regional ice-flow indicator and clast provenance studies in the Deer Lake basin area, in an attempt to resolve conflicting interpretations of the glacial history presented by Rogerson (1979) and Vanderveer and Sparkes (1982). In coastal areas, Brookes (1974) described the deglacial history of the Bay of Islands, and identified a raised marine delta at the mouth of the Humber River near Corner Brook (Plate 1). No radiocarbon dates are available for the delta, although a date on a delta at a similar elevation at Cox's Cove was used to infer a date of about 12,600 years BP for the Humbermouth delta (Brookes, 1974). Batterson and Taylor (1990) speculated on the existence of a large proglacial lake in the Deer Lake basin during deglaciation, followed by a period of marine incursion as suggested by Batterson and Kirby (1988). The relationship of this period of marine incursion to the timing of the Humbermouth delta development is uncertain.

FIELD PROGRAM

Bedrock outcrops were examined for striae and other ice-flow indicators. Fifty-nine new striation sites were recorded to supplement those identified during previous projects. Detailed descriptions of surficial sediment were made from natural or man-made exposures, including 25 backhoe test-pits. Sedimentological properties examined included texture, compaction, sedimentary structures and clast lithology. Clast fabrics were taken from 56 diamicton exposures. In each case, 25 elongate pebbles having a length:breadth ratio of greater than 3:2 were measured. Results were plotted on a stereogram and analyzed using the Stereo™ software package for the Apple Macintosh microcomputer (MacEachran, 1990). Principal eigenvalues, which measure the strength of the fabric orientation, and K values, which measure the distribution of the clast orientations (i.e., plunges), were produced using the method outlined by Woodcock (1977). Normalized eigenvalues (S_1) can range between 0.33 (random) and 1.0 (unidirectional). The K values of less than 1.0 suggest girdle distributions. In this paper, strong fabrics are defined as those with $S_1 > 0.6$ and $K > 1.0$.

Matrix samples were taken from 156 locations for textural and possible geochemical analysis. Matrix samples (finer than 2 mm/-1ϕ) are sieved through a nest of 6 stainless-steel sieves (-1ϕ to 4ϕ), and the silt–clay fraction (finer than 4ϕ) analyzed for grain-size distribution using a Coulter Counter (Model TAII-L, 100 and 200 mm apertures). Grain-size results are not currently available. Clast samples (coarser than 64 mm/-6ϕ) of between 50 and 100 clasts were taken from 126 locations, and the rock types were identified.

ICE FLOW

Striation sites having more than one direction of flow are rare and generally confined to the Humber River valley (Figure 3). The soft Carboniferous sediments in the area rarely permit the preservation of striae, but the few striae found showed an early flow generally parallel to the valley crosscut by a later regional flow across the valley. Striae related to the early flow were found within the Humber River valley as far downstream as the eastern end of the Humber gorge (Plate 2). The direction of the later regional ice-flow event is variable across the field area. South of Corner Brook, ice flow is northeastward (310° ± 10°). North of the Steady Brook valley, ice flow is westward (270° ± 10°), and parallels major outlet valleys, e.g., Old Mans Pond (Plate 3). In the Corner Brook area, ice flow was into the Humber Arm from all sides, suggesting it was a major channel for ice.

Well-oriented clast fabrics in glacial diamictons (i.e., $S_1 > 0.6$ and $K > 1.0$) can also indicate ice movement (Harrison, 1957; Lawson, 1981; Dowdeswell and Sharp, 1986). Where found, these fabric orientations are generally consistent with the erosional evidence.

The ice-flow data suggests an early topographically controlled flow down the Humber River valley from a source to the north, possibly the Long Range mountains, as suggested by Vanderveer and Sparkes (1982). Clasts from the Long

Range, however, were not recognized. This early flow was followed by a regionally consistent ice-flow event from a source to the east of Grand Lake. Toward the coast, ice flow was affected by the orientation of major valleys.

CLAST PROVENANCE

The ice-flow pattern from striae provides data on the general direction of glacial flow, but not on the location of ice dispersal centres. Rock types identified from glacial diamictons can provide data on both direction and distance of glacial transport, if those rock types are from a discrete source area. Clast fractions in diamictons in the field area were dominated by local rock types, commonly metasediments or Carboniferous sediments. Exotic clasts commonly comprised less than 10 percent of clasts collected, and most of these had wide potential source areas. However, several rock types are diagnostic of particular provenance areas. Red, flow-banded rhyolite and silicic tuff clasts in the diamictons were probably derived from the Springdale Group. A red, medium- to coarse-grained one-feldspar-granite, and quartz–feldspar porphyry clasts were derived from the Topsails intrusive suite. Pink, medium- to coarse-grained K-feldspar porphyritic two-feldspar-granite clasts were eroded from the Hinds Lake granite. When found in association with each other, these clasts indicate an ice-dispersal centre on the Topsail Hills, centred on the Hinds Lake area. Clasts originating from the Buchans Group in the Red Indian Lake region are not found in the field area, suggesting that any ice centre was located to the west of Red Indian Lake.

QUATERNARY SEDIMENTS AND FEATURES

A wide range of sediment types were found including tills, subaqueous fan deposits, lacustrine, marine and fluvial sediments. Brief descriptions of these sediments follow.

DIAMICTON

Diamicton is common across the whole area, except within the Humber River valley (Figure 4). The character of diamictons is mainly influenced by the underlying bedrock geology. On the uplands, underlain by metasediments to the south of Corner Brook, diamictons are thin and discontinuous. They have a Munsell colour of light olive brown (2.5Y 5/4) to dark yellowish-brown (10YR 4/4) when moist, and light grey (2.5Y 7/2) when dry, and have a dominantly sandy matrix with less than 20 percent silt and clay. Clasts are mostly granules to boulders of local provenance, with the largest clasts up to 60 cm diameter, subrounded to subangular, rarely striated, and commonly have a weak to moderate clast fabrics. Exotic clasts account for less than 5 percent of clasts and are commonly derived from the Topsail Hills. In section, these diamictons commonly appear structureless, although thin (< 2 mm), irregular-shaped sorted lenses are common beneath clasts.

Diamictons, on sedimentary bedrock on the uplands around Old Mans Pond, are generally thin and discontinuous. They have a Munsell colour of olive (5Y 5/3) to dark greyish-brown (2.5Y 4/2) when moist, and light yellowish-brown (2.5Y 6/4) to pale yellow (5Y 7/3) when dry, and have silty sand matrices with 20 to 50 percent silt and clay. Clast characteristics are similar to those described for the Humber River valley, with clasts commonly originating from the Topsail Hills.

Those diamictons in areas dominated by Carboniferous bedrock have a silty sand matrix with 20 to 45 percent silt and clay, and have Munsell colours of grey brown (2.5Y 5/2) to reddish-brown (5YR 4/4) when moist, and very pale brown (10YR 7/3) to reddish-yellow (7.5YR 6/6) when dry. The clast content is dominated by local rock types, although clasts from the Topsail Hills are common.

Generally, diamictons derived from Carboniferous bedrock are thicker than in other areas, and in several places, more than one diamicton was exposed at a single site. In each case, the internal character of the diamicton units is different, and therefore, the regional significance of diamicton distribution cannot currently be assessed. Typical is a section, 100 m northeast of the Pasadena dump, where three different diamictons are exposed (Plate 4). The lower diamicton is dark brownish-grey (10YR 4/2, moist) and structureless. Clast fabrics are aligned parallel to the regional ice-flow direction and are strong (S_1 = 0.79 and 0.92 and K = 1.21 and 1.61). The middle diamicton is olive grey (5Y 4/2, moist) having a silty sand matrix and 38 percent fines. The unit has a blocky structure. Clasts are subangular to subrounded, commonly striated and of predominantly local origin. Small, irregular-shaped lenses of medium to coarse sand are common beneath clasts. Clast fabrics are aligned with the regional ice flow and are strong (S_1 = 0.73 and 0.72 and K = 5.32 and 8.40). The upper diamicton is reddish-brown (5YR 4/2, moist) having similar characteristics to the underlying unit. Clast fabrics are similar as immediately above and are moderate to strong (S_1 = 0.69 and 0.74 on 2 fabrics) with inconsistent K-values (0.51 and 3.65). The diamictons become more consolidated with depth, although grain size remains fairly uniform.

The characteristics of the diamictons described above and those found elsewhere are generally consistent with a subglacial origin, possibly meltout till (Lawson, 1981; Haldorsen and Shaw, 1982; Shaw, 1982; Dowdeswell and Sharp, 1986; Dreimanis, 1988). The striated clasts, their subrounded form and well-oriented clast fabrics are consistent with basal transport. The presence of sorted layers beneath clasts suggests the presence of water during deposition. Clast fabrics are variable across the field area. Using the methods of Mark (1974) and Woodcock (1977), fabrics were plotted on the graph illustrated in Figure 5. Fabrics with a strong cluster plot on the top left of the graph, and those with girdle fabrics plot on the lower right. Rappol (1985) suggests that fabrics that cluster are consistent with undisturbed basal meltout and lodgement tills, whereas those that have girdle distributions are consistent with disturbed basal tills and debris flows. Figure 5 shows that disturbed and undisturbed sediments are found in the field area. Most of the fabrics

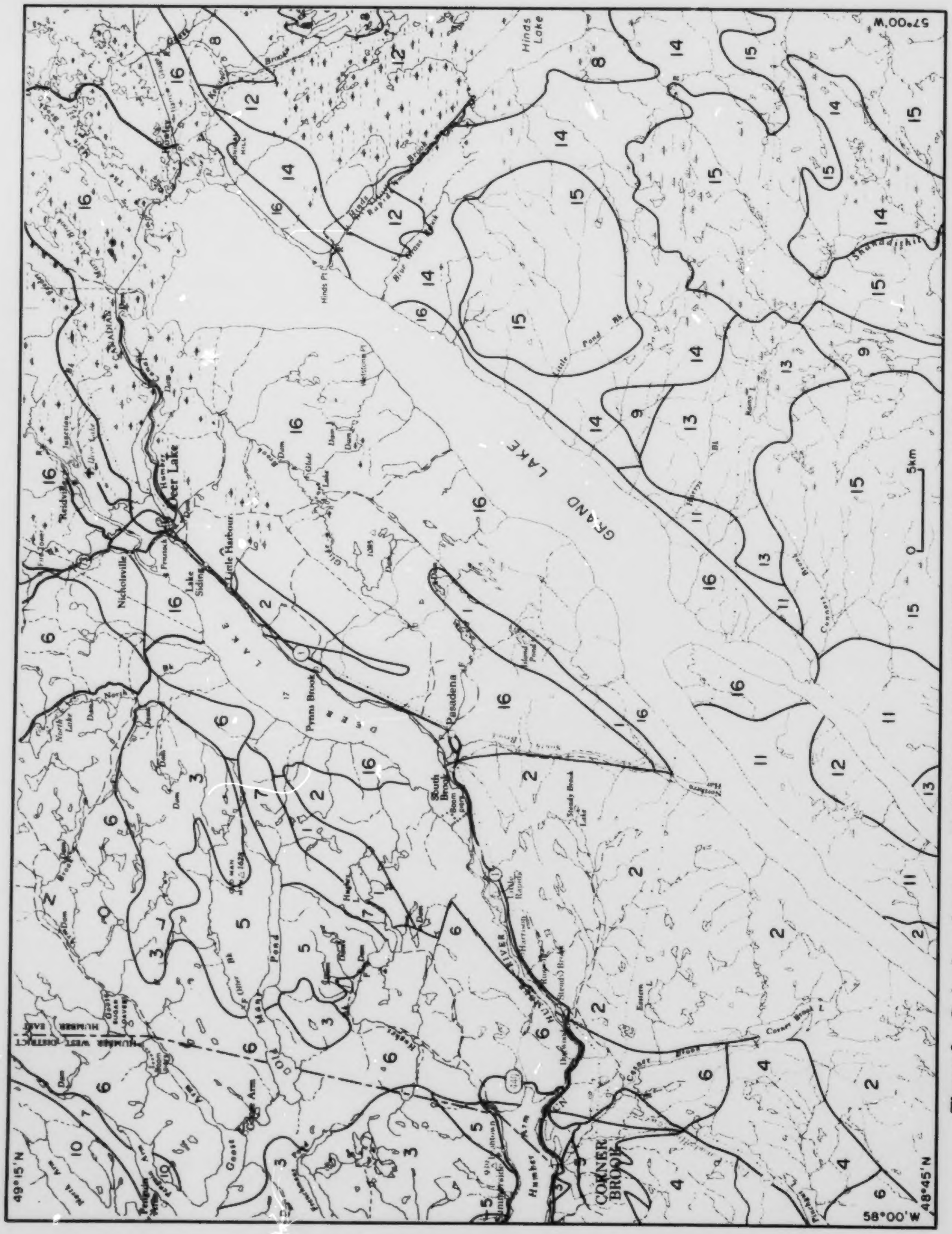
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Figure 2. *Bedrock geology map. Modified from Hyde (1984), Whalen and Currie (1988) and Williams and Cawood (1989).*

LEGEND (for Figure 2)

CARBONIFEROUS

16 *Carboniferous clastic sedimentary rocks*

SILURIAN

15 Topsails Intrusive Suite: *rhyolite, one-feldspar and two-feldspar granites; some have peralkaline affinities*
14 Springdale Group: *flow-banded rhyolite, tuff and basalt*
13 Rainy Lake Complex: *gabbro, diorite and granodiorite*

ORDOVICIAN

12 *Massive to slightly foliated granite and granodiorite*
11 Glover Group: *basalt, diabase, tuff and conglomerate*
10 *Grey to black scaly shale*
9 *Massive to moderately foliated granodiorite and minor tonalite*
8 Hungry Mountain Complex: *moderately to strongly foliated gabbro to granite*

CAMBRIAN

7 *Grey to black shale and light grey platy limestone*
6 *Shale, limestone and dolostone (includes Reluctant Head Formation)*
5 *Grey to black shale and thin white quartzite*
4 *Crystalline limestone and phyllite*

UPPER PROTEROZOIC TO LOWER CAMBRIAN

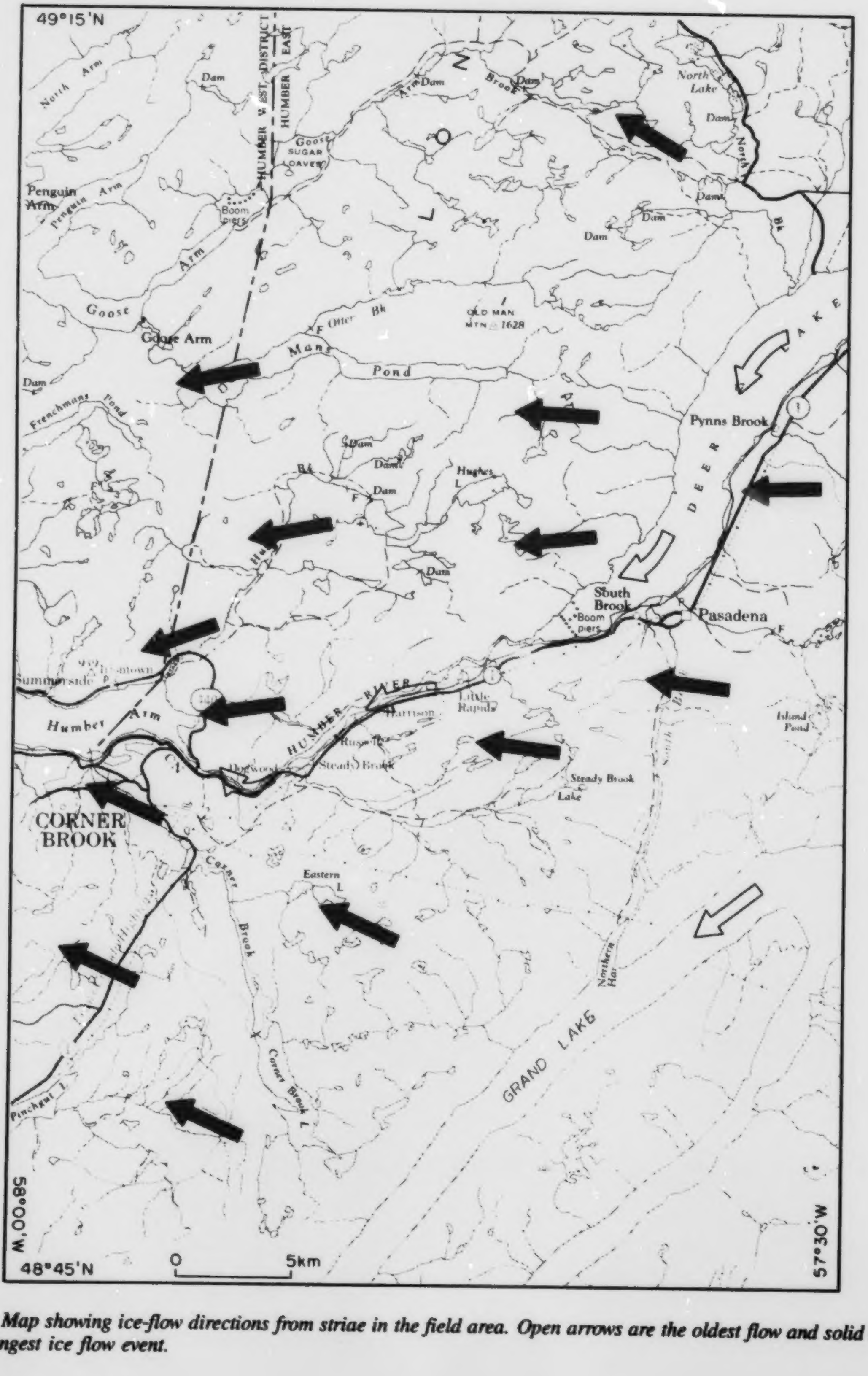
3 *Greywacke, quartzite*
2 Mount Musgrave Group: *psammitic and pelitic schist*
1 Hughes Lake Complex: *metavolcanic rocks and granite*

Plate 1. *Raised marine delta at the eastern end of the Humber Arm near Corner Brook. The delta has a surface elevation of about 51 m asl and was probably formed about 12,600 years BP.*

indicative of disturbed sediments are found in the Wild Cove valley and other areas of steep slopes. Fabrics from diamictons in the Pasadena dump section plot mostly in the undisturbed basal till area. In lodgement tills, well-oriented clast fabrics result from friction with the substrate, whereas in meltout tills, clast fabric is inherited from the glacial transport process (Dreimanis, 1988). In both cases, the resultant fabric is commonly parallel to ice movement, although Lawson (1981) suggests that resedimentation during and following deposition may weaken clast fabrics. Evidence was not found to suggest that the diamicton units exposed at the Pasadena dump site or elsewhere represent more than one glacial episode.

GLACIOFLUVIAL AND FLUVIAL SEDIMENTS

Glaciofluvial and postglacial fluvial sediments are common within the Humber River valley, and in isolated areas within the South Brook valley and near Old Mans Pond (Figure 4). However, exposures through this sediment are relatively rare, being confined to small deposits in the South Brook valley and at the western end of the Humber gorge. Little detailed sedimentological analysis has yet been completed on these sediments and, therefore, only a brief


49°15'N
48°45'N
58°00'W
57°30'W
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North Arm
Penguin Arm
Goose Arm
Goose SUGAR LOAVES
Boom piers
HUMBER WEST DISTRICT
HUMBER EAST
North Lake
North Bk
OLD MAN MTN 1628
Otter Bk
Mans Pond
Frenchmans Pond
DEER LAKE
Pynns Brook
Hughes L
South Brook
Pasadena
Summerside
Humber Arm
HUMBER RIVER
Little Rapids
Harrison
Steady Brook
Steady Brook Lake
Island Pond
CORNER BROOK
Corner Brook
Eastern L
Corner Brook L
Pinchgut L
Northern Har
GRAND LAKE
Dam

Figure 3. *Map showing ice-flow directions from striae in the field area. Open arrows are the oldest flow and solid arrows are the youngest ice flow event.*

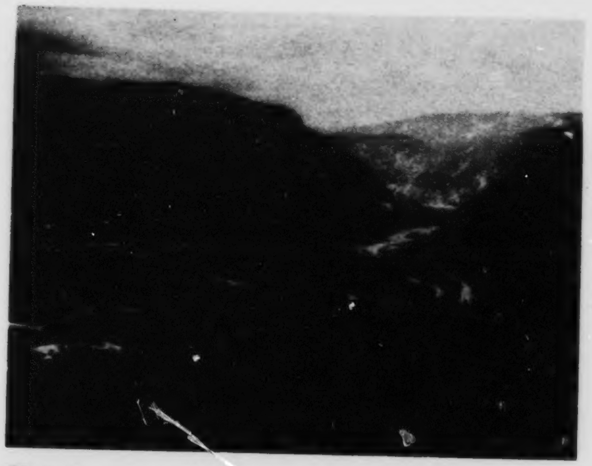

Plate 2. *View of Humber gorge near Corner Brook. The valley has evidence of ice flow through it, although most flow was likely through the Wild Cove valley to the north. Ice remaining in the gorge during deglaciation may have impounded a large proglacial lake and following its drainage was inundated by the sea.*

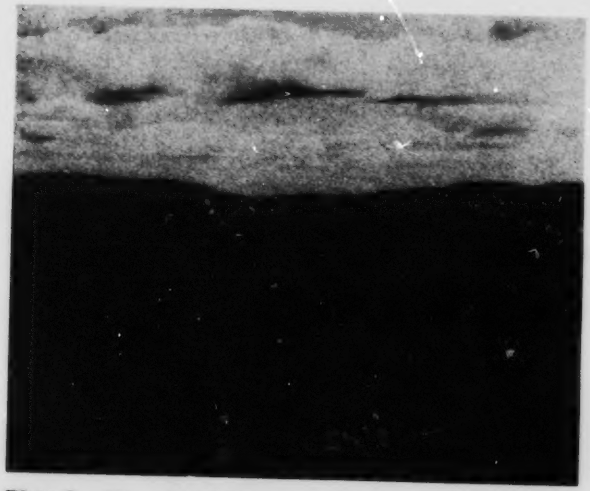

Plate 3. *View of Old Mans Pond looking east. This valley was a major outlet of ice from the Topsail Hills toward the coast. Sediments on the northern valley slopes are largely colluviated tills.*

overview is presented here. In Dawe's Pit, at the western end of the Humber gorge, interbedded sands and pebbly sands are exposed. The sand units are moderately to well-sorted planar beds, with some planar tabular crossbedding, indicating westward flow. Pebbly sands are roughly planar bedded, and contain subrounded clasts of mixed rock types. Normal faults are common throughout the unit, with vertical displacements of up to 15 cm. Faulting is likely the result of the collapse or settling of sediment due to the melting *in situ* of buried ice incorporated during deposition, and is common in glaciofluvial sediments (Sugden and John, 1976). Fluvial sediments commonly do not contain collapse structures such as those exposed in Dawe's Pit.

In drill core recovered from the Steady Brook area by the Department of Works, Services and Transportation (D. Brennan, personal communication, 1990), gravels are found near the base of several 50 to 60 m cores. Their stratigraphic position, underlying silts and clays of probable marine origin (as discussed further on) suggests that these sediments are glaciofluvial in origin.

Postglacial fluvial gravels are found stratigraphically above the marine muds. They were well exposed during the summer of 1991 due to road construction within the Humber gorge near Corner Brook. A typical stratigraphy consisted of a basal unit of planar bedded, moderately to well-sorted, medium to fine sand, with occasional ripples with a 10 to 15 cm wavelength and 1.5 cm amplitude. These sands are overlain by 50 to 100 cm of planar-bedded, clast-supported pebble gravel. Clasts are subrounded, of mixed rock types and have a 2- to 3-mm-surface coating of reddish-brown silty clay. Overlying this gravel unit are poorly sorted, planar-bedded sandy gravels that have a fine to medium sand matrix. The unit commonly contains ovoid, open-work gravel lenses. The stratigraphy suggests increasing current flow, possibly related to decreasing water depths within the Humber gorge associated with falling sea levels. To the east of the Humber gorge gravelly sediments are uncommon. More usual are moderately sorted, fine to medium, planar cross-stratified sands. These suggest deposition under more quiescent flow conditions.

SEDIMENTS AND FEATURES ASSOCIATED WITH HIGHER LAKE LEVELS

Batterson and Taylor (1990) suggested the possibility of a large proglacial lake covering the modern Deer Lake–Grand Lake–Sandy Lake–Birchy Lake basins. This was based on evidence of fan deltas described in the Pasadena, Grand Lake and Birchy Lake areas, all of which had similar elevations (Lundqvist, 1965; Liverman and St. Croix, 1989). Within the field area, Batterson and Taylor (1990) identified a single delta at 135 m asl in the South Brook valley, 3 km south of Pasadena. During this field season, two additional deltas were found in the South Brook valley. One was found about 1.8 km south of the previously documented site adjacent to a Hydro transmission line at a surface elevation of 135 ± 2 m asl (altimeter estimate). The second delta was at Northern Harbour on Grand Lake, at 130 m ± 2 asl (43 m above Grand Lake, altimeter estimate). Sediments associated with the proglacial lake could not be positively designated, although massive clayey silts found at several locations near the modern South Brook valley possibly represent lacustrine deposition. However, an alluvial origin for these sediments could not be discounted.

SEDIMENTS AND FEATURES ASSOCIATED WITH HIGHER SEA LEVELS

Coastal features associated with higher sea-level stands have been documented in the Bay of Islands (e.g., Flint, 1940; Brookes, 1974). Within the field area, Brookes (1974) described a prominent marine delta at Humbermouth and

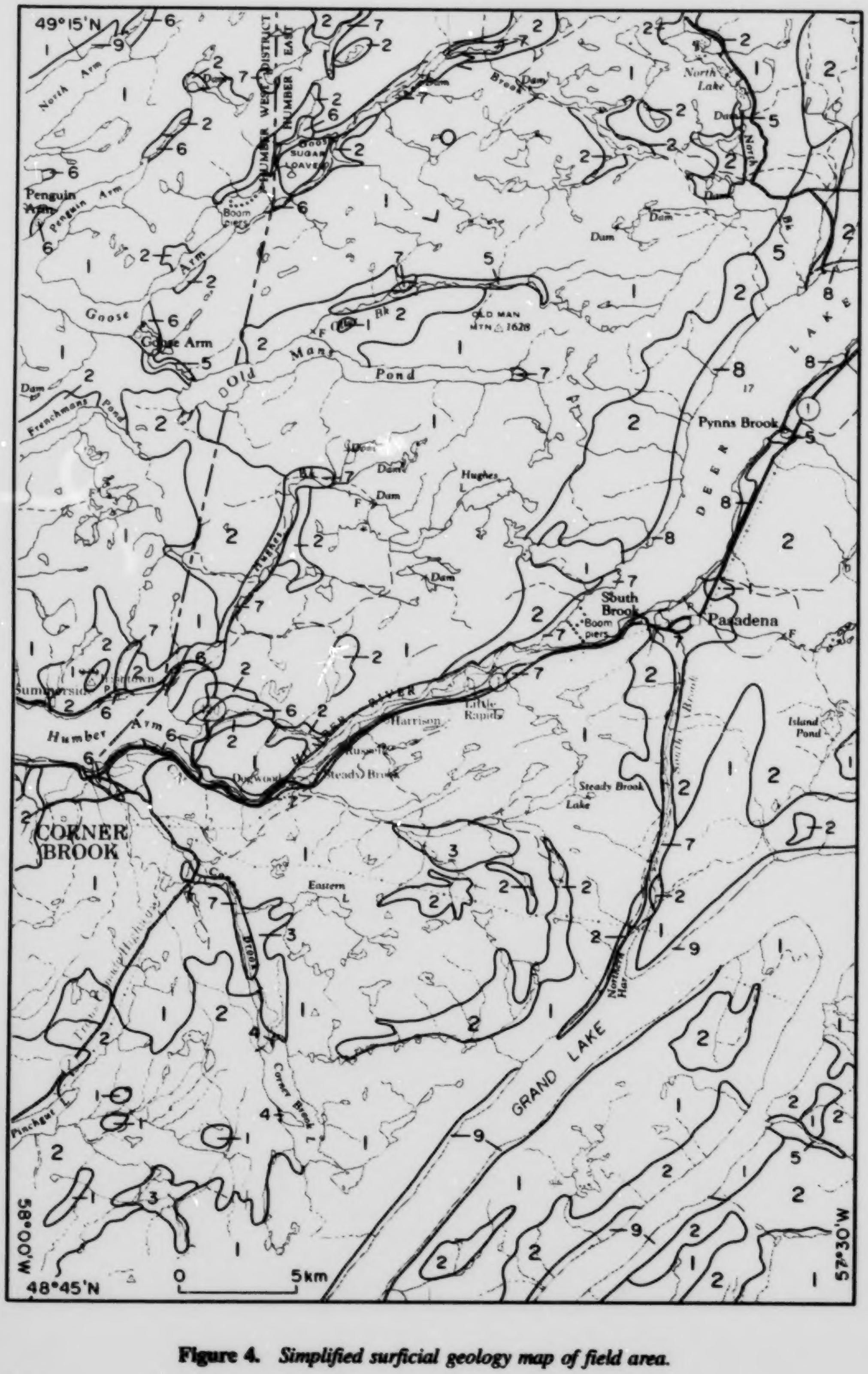
49°15'N
North Arm
Penguin Arm
HUMBER WEST DISTRICT
HUMBER EAST
GOOSE SUGAR LOAVES
Boom piers
North Lake
Goose Arm
OLD MAN MTN 1628
Old Man Pond
Frenchmans Pond
Pynns Brook
DEER LAKE
Hughes L
South Brook
Boom piers
Pasadena
Summerside
Humber Arm
HUMBER RIVER
Little Rapids
Harrison
Island Pond
Dogwood
Steady Brook
Steady Brook Lake
CORNER BROOK
Eastern L
Corner Brook
GRAND LAKE
Pinchgut
58°00'W
48°45'N
0
5km
57°30'W

Figure 4. *Simplified surficial geology map of field area.*

LEGEND (for Figure 4)

POSTGLACIAL

9 Colluvium: *material derived from adjacent slopes and gravity deposited*

8 Lacustrine: *sediment 1 to 10 m thick composed of bedded sands and gravels; includes beach ridges and terraces. Common along shores of Deer Lake*

7 Fluvial: *sediment 1 to 10 m thick, composed of planar-bedded, moderately to well-sorted sands and gravels. Especially common in Humber River and Hughes Brook valleys*

6 Marine: *sediment 1 to 30 m thick composed of poorly to well-sorted sands and gravels, and rhythmically bedded silts and clays; includes marine terraces and deltas*

GLACIAL

5 Glaciofluvial deposits: *sediment 1 to 30 m thick composed of poorly to well-sorted sands and gravels*

4 Till ridges, transverse to flow: *ridges composed of subglacial diamicton; relief 2 to 5 m*

3 Till hummocks: *mounds of basal diamicton having no consistent orientation; relief 2 to 5 m*

2 Till veneer: *(< 2m thick): Sediment generally of local provenance. Numerous bedrock exposures throughout unit*

PREGLACIAL

1 Bedrock: *includes exposed bedrock and bedrock having a thin (<1m) vegetation or sediment cover*

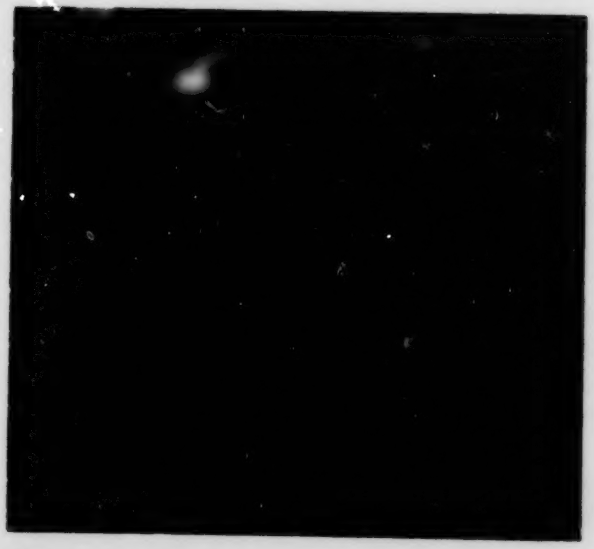

Plate 4. *Three diamictons exposed near the Pasadena dump. The diamictons are distinguishable on the basis of colour and compaction, although textural, fabrics and clast provenance are similar between all three units.*

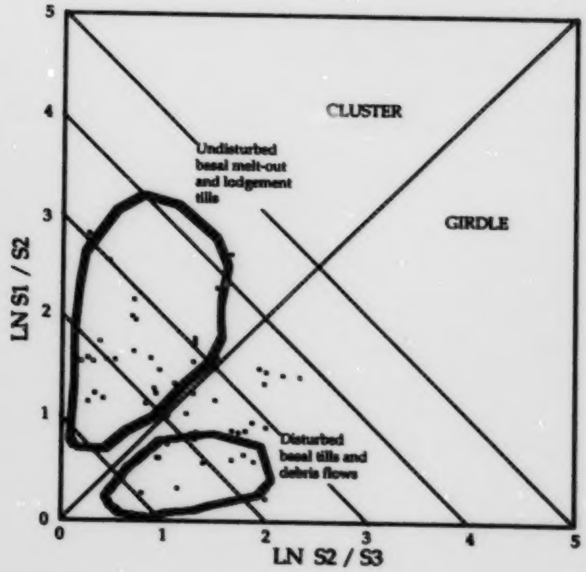

Figure 5. *Graph of fabric data following the methods of Mark (1974), Woodcock (1977) and Rappol (1985). The graph illustrates potential depositional environments for diamictons in the field area.*

assigned an age of about 12,600 years BP based on shell dates from a delta of similar elevation at Cox's Cove. The surface of this delta is 51 m asl (altimeter estimate). Features at similar elevations were found along the coast near Hughes Brook and at Irishtown. Inland deltas were identified at the head of the Wild Cove valley at about 53 m asl (altimeter estimate) and in the lower Hughes Brook valley at an elevation of about 59 m asl (altimeter estimate). The relationship of these delta surfaces to sea level is currently not clear. Within the Humber River valley several deltas have been identified. At Little Rapids a fan delta is located at about 45 m asl (altimeter estimate). The delta on which the community of Pasadena is situated has a surface elevation of 33 m asl (altimeter estimate), as is a delta at Pynn's Brook (Plate 5), and a delta near Coal Brook about 5 km southwest of Nicholsville is at 47 m asl, similar to the surface elevation of a delta at Nicholsville at 48 m asl. Several terraces have also been noted on both sides of Deer Lake (Batterson and Kirby, 1988). Precise elevations of those on the north side of Deer Lake are currently unknown, but 2 terraces on the south side have bases at 21 and 36 m asl (altimeter estimates).

Plate 5. *Raised marine delta at Pynn's Brook, east of Pasadena. The delta has a surface elevation of 33 m asl, and was formed during a higher sea level in the modern Deer Lake basin about 12,000 to 12,500 years BP.*

Fossiliferous silty sands found in the Humber gorge suggest that many of the features at elevations below 50 m asl are likely of marine origin. The shells are *Balanus hameri*, a barnacle whose preferred environment is water deeper than 20 m, 0 to 5°C summer surface-water temperatures, and salinities in excess of 33‰ (Pilsbury, 1916). This environment would not occur adjacent to a melting ice mass, and, therefore, marine incursion through an open Humber gorge and into the Humber Valley is indicated. Terraces and deltas found at various elevations are likely related to stillstands in sea-level lowering. Sediments related to higher sea-level stands have been identified on the modern coast. Batterson and Kirby (1988) report fossiliferous silts and clays at Cooks Brook, west of Corner Brook, and shells have also been noted in the rhythmically bedded silts and clays at Wild Cove (M.J. Ricketts, personal communication, 1991). Similar rhythmically bedded silts and clays have been described from many locations in the Humber River valley (Batterson and Kirby, 1988). Although their origin is still uncertain, it is likely they have a marine origin because of their similarity to sediments found in coastal areas, and their stratigraphic position below the sandy to gravelly postglacial fluvial sediments described above.

QUATERNARY HISTORY–PRELIMINARY OBSERVATIONS

Initially, ice flow in the field area was topographically controlled. Striation evidence suggests early movement of valley glaciers down the Deer Lake and Grand Lake basins. Ice in Deer Lake probably exited to the coast through Wild Cove, although minor flow through the Humber gorge is possible. The source of this ice is unknown, although a northerly source is likely based on the striation patterns. As the ice masses developed, topography had less of an influence, and dispersal centres shifted to the Topsail Hills as indicated by striations and the wide dispersal of clasts identified as originating from the Topsail Hills. Diamictons are common across the area, although they are thin over the metasediment-dominated uplands to the south and east of Corner Brook. Thicker diamictons are found in areas underlain by Carboniferous sediments. Most glacial diamictons identified were basal tills.

In terms of surficial deposits and landforms, much of the Corner Brook–Pasadena landscape was formed by events in the immediate postglacial period. During deglaciation, when ice retreated to the highlands, a proglacial lake developed in the Deer Lake basin, and possibly in surrounding areas. Evidence for this lake within the field area is several deltas at about 130 to 136 m asl. Diamictons with girdle fabrics (Figure 4) may have been reworked during periods when they were submerged by higher lake or sea levels. Probable fans were identified on aerial photographs. These were not extensively examined in the field, but occur in the Wild Cove valley below maximum sea-level stand, and in the South Brook valley below the inferred surface of the proglacial lake described earlier. Following drainage of this lake, which was likely through the Humber gorge, marine incursion flooded the Humber River and Hughes Brook basins. As sea level fell, a series of terraces and deltas was developed in the Humber River valley and along Hughes Brook. Marine sediments were probably eroded from some areas, such as the Humber gorge, where mostly coarse-grained postglacial fluvial sediments are found. To the east of the gorge and in the Hughes Brook valley, fluvial sediments are generally finer grained, suggesting a lower energy fluvial system.

IMPLICATIONS FOR LAND-USE PLANNING

The wide range of sediment types has significant implications for land-use planning within the area. The marine silts and clays are impermeable, and therefore, downward percolation of water and/or pollutants is severely restricted. Adequate drainage of overlying sediment is essential to prevent destabilization of surface slopes. Slope failure is a potential problem in steeply sloped areas underlain by these muds. Similarly, selecting waste-disposal sites requires consideration of the underlying surficial geology to ensure that the rapid transport of effluents into local streams and groundwater systems does not occur. The current site of the Corner Brook dump at Wild Cove, which is underlain by marine clays, should be monitored to ensure that potential pollution levels in the Bay of Islands are controlled. The

Humber Valley below about 50 m asl is likely underlain by marine muds. These sediments are commonly encountered in excavations in Pasadena and other communities in the Humber River valley. They will likely be a significant engineering problem during the proposed reconstruction of the Trans Canada Highway between Corner Brook and Deer Lake. More work is required to define the distribution and physical properties of these marine sediments.

In areas south of Corner Brook, the surficial cover is thin and soil development generally poor. The rapid denudation of the forest cover is resulting in soil erosion that could ultimately result in lower potential for revegetation of these areas. Monitoring of erosion rates in recently deforested areas is recommended to ensure erosion does not proceed unabated.

ACKNOWLEDGMENTS

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BEDROCK-AGGREGATE ASSESSMENT OF THE NORTHERN PENINSULA OF NEWFOUNDLAND

D. Bragg
Terrain Sciences Section

ABSTRACT

The 1991 field season consisted of an evaluation of bedrock-aggregate potential of the Northern Peninsula. This is a continuation of a reconnaissance aggregate assessment started in 1985, to determine the bedrock-aggregate potential of the different rock groups and formations in the region.

A total of 368 sites were visited and 233 samples collected. A petrographic number was assigned to each sample (based on preliminary petrographic examination). A total of 127 samples are considered to be of high quality, 34 samples are considered to be of marginal quality and 72 samples are considered to be of low quality based on their petrographic numbers.

Further analysis such as Los Angles Abrasion, Magnesium Sulphate Soundness and Alkali-Aggregate Reactivity testing are planned for selected samples.

INTRODUCTION

During the 1991 field season, an evaluation of bedrock-aggregate potential from Deer Lake to St. Anthony on the Northern Peninsula (Figure 1), excluding Gros Morne National Park, was undertaken. This assessment was conducted in order to determine the quantity and quality of bedrock aggregate in the area for local industrial use, such as road construction and bridge repair or replacement. This inventory should help to eliminate the use of poor-quality aggregate such as alkali-reactive rock that reacts with cement to cause premature destruction of concrete, and friable rock that has been used as a form of rip-rap or armour stone, or in asphalt mix.

An additional objective was to locate potential sites for further detailed investigation in order to find a suitable replacement for sand, which is presently being transported from Deer Lake.

FIELD WORK

Field work consisted of a detailed examination of all rock quarries, road-cuts and natural outcrops along all highways and side roads in the study region. Each site investigation entailed rock identification, representative sampling, the recording of all geological features present, and the determination of petrographic number, following the methods outlined by Bragg (1989).

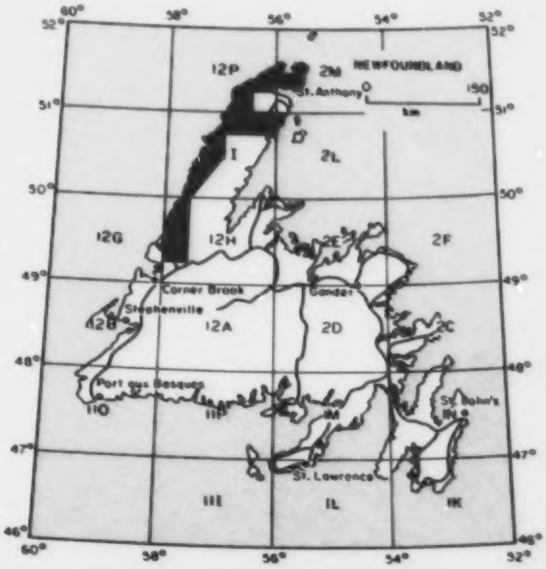

Figure 1. *Location of study area.*

At least one representative sample (1 kg) was collected from all quarry sites and selected samples were collected from road-cuts or natural outcrops if the rock types were consistent for these sites. However, between 2 and 5 samples were commonly taken if the site was larger than 400 m², if there were a number of different rock types present, if the degree of weathering varied, or if veining and alteration zones were observed. Much larger samples (50 kg) were taken at selected sites based on rock type, quantity and freshness of the sample for more extensive laboratory analyses.

GENERAL GEOLOGY

The geology of the study area, which is briefly summarized below, has been studied in detail by Schuchert and Dunbar (1934); Knight and James (1987); Chow (1985); Cumming (1983); Knight and Boyce (1987); Kindle and Whittington (1958); James and Stevens (1986); Klappa *et al.* (1980); Stenzel *et al.* (1990); Cooper (1937); Phair (1949); Williams and Smyth (1983); Stouge (1982) and Cooper (1937).

The St. George Group consists of six formations, but, only four formations, the Catoche (rubbley, blue-grey, bioturbated, micritic limestone), Boat Harbour (biomicritic and stromatolitic limestone and laminated dolostone), Watts Bight (dark-grey to black, cherty, crystalline dolomite having large stromatolite mounds), and the Southern Arm (bioturbated micritic limestone, minor dolostone and abundant black chert) are found in the field area. A total of 52 samples were collected from this group.

The Port Au Port Group consists of two formations; the Petit Jardin Formation comprises shale, limestone and dolostone, which Levesque (1977) divided into three informal members, and the March Point Formation, which consists of grey, bioturbated limestone and shale. A total of 53 samples were collected from the Port Au Port Group.

The Labrador Group consists of five formations, however, only three formations, the Hawke Bay Formation (white quartzite, minor shale, sandy dolomite, and oolitic limestone), the Forteau Formation (shale, calcareous siltstone, minor limestone, oolitic limestone and quartz sandstone), and the Bradore Formation (red to buff arkosic sandstone, minor conglomerate and hematitic sandstone) were sampled in the field area. A total of 36 samples were collected from this group.

The Cow Head Group, which is undivided, consists of interbedded limestone breccia, lime mudstone, calcarenite, red, green and purple shale and chert. Only 4 samples were collected from this group. The Table Head Group, which is undivided in the field area, consists of limestones, shales, ribbon limestones, breccias, sandstones and siltstones. A total of 15 samples were collected. The Goose Tickle Group, also undivided, consists of greywacke and slate; only 1 sample was collected from this group.

The Maiden Point Formation consists of coarse-grained greywacke, red and green slate, quartz-pebble conglomerate, massive basalt flows, mafic agglomerate, tuff, minor pillow lava and medium- to coarse-grained diorite and gabbro. A total of 43 samples were collected from this formation.

The Longe Range Complex consists of quartzofeldspathic gneiss and schist, psammitic to pelitic schist, amphibolite and granite; a total of 3 samples were taken from this area.

The Goose Cove Formation consists of epidote, amphibolite and greenschist facies, metavolcanic rocks and minor metasedimentary rocks; psammite and pelite are at the base of the formation. A total of 11 samples were taken from this area.

The Epine Cadoret Formation consists of greywacke and brown-weathering slate, and minor pebble conglomerate. A total of 4 samples were collected from this formation.

The Northwest Arm Formation consists of black, green and grey shale with calcareous siltstone beds and limestone; only 1 sample was collected. Melange-type material consists of ultramafic and sandstone blocks in a black shale–slate matrix; a total of 10 samples were collected.

RESULTS

Table 1 shows the initial assessment of the quality of the different rock groups–formations based on field observations and petrographic numbers. Petrographic factors (Canadian Standards Association, 1973) are assigned to different rock types with 1 (best), 3 (fair), 6 (poor) and (10 deleterious). The St. George, Port Au Port and Table Head groups and the Long Range Complex are all considered to be of high-quality aggregate potential and should be investigated further. The majority of samples (123) come from these units, and 99 samples were considered to be of high potential (P.N. 105 to 150) and 24 samples were considered to be of marginal quality (P.N. 160 to 245). There were no poor-quality samples taken from these areas. The Cow Head Group, Epine Cadoret Formation and Goose Tickle Group are considered to be of marginal quality, with sporadic occurrences of high-quality material. The Labrador Group, Maiden Point and Goose Cove formations range from high- to poor-quality aggregate (P.N. 110–625). The melange unit is considered to be of poor quality (P.N. 310–450) for construction aggregate.

Table 1. Petrographic numbers for selected areas on the Northern Peninsula (Deer Lake to St. Anthony)

Group/Formation	# of Samples	Petro. # Range	Petro. #<150	Petro. #>150
St. George Gp.	1	110	1	0
Catoche Fm.	11	115-325	10	1
Boat Harbour Fm.	15	110-155	15	0
Watt s Bight Fm.	20	110-145	20	0
Southern Arm Fm.	5	110	1	0
Port au Port Gp.	0	0	0	0
Petit Jardin Fm.	33	125-175	23	10
March Point Fm.	20	110-185	13	7
Labrador Gp.	8	110-625	5	3
Hawke Bay Fm.	3	120-250	2	1
Forteau Fm.	23	115-455	17	6
Bradore Fm.	2	180	0	2
Cow Head Gp.	4	185-255	0	4
Table Head Gp.	15	110-165	13	2
Long Range Cpx.	3	120	3	0
Melange	10	310-450	0	10
Maiden Point Fm.	43	115-450	9	34
Epine Cadoret Fm.	4	185-210	0	4
Northwest Arm Fm.	1	110	1	0
Goose Tickle Gp.	1	355	0	1
Goose Cove Fm.	11	125-265	3	8

SUMMARY

The carbonate rocks of the St. George, Port Au Port, and Table Head groups and the igneous rocks of the Long Range Complex have the highest potential for use as high-quality aggregate. The sedimentary rocks of the Labrador Group, Maiden Point Formation and Goose Cove Formation range from high to poor quality and further study or investigation such as Los Angeles Abrasion, Magnesium Sulphate Soundness and Alkali-Reactivity testing are recommended on these rocks. The melange unit, which is considered to be of poor quality (P.N. 310—450), should not be considered for aggregate use.

ACKNOWLEDGMENTS

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BEDROCK-AGGREGATE POTENTIAL OF THE GREAT MOSQUITO COVE AREA, BULL ARM

D. Bragg and C.F. O'Driscoll[1]
Terrain Sciences Section

ABSTRACT

The bedrock geology of the Great Mosquito Cove area was mapped to determine the quantity and quality of bedrock for potential use in the construction of the Gravity-Based Structure for the Hibernia project. The study consisted of geological mapping and detailed testing of the selected rock units found in the area. Geotechnical bedrock properties, including geological structures and deleterious substances were examined; abrasion and soundness testing were analyzed, petrographic analyses and petrographic number computations were carried out. Results show that there are sufficient quantities of high-quality bedrock in the area for use as aggregate.

INTRODUCTION

The Great Mosquito Cove area is situated on the western side of Bull Arm, an inlet that extends northwestward from Trinity Bay on the Isthmus of Avalon. The closest community is Sunnyside, which lies at the northwestern extremity of Bull Arm. A recently constructed road connects the area to the Trans-Canada Highway, 6 km to the west.

The production system for the Hibernia oilfield includes a Gravity-Based Structure (GBS) and a Topsides module. Construction of the GBS, Topsides module, and mating of the Topsides with the GBS will take place at Great Mosquito Cove. The site is presently being prepared for construction.

High-quality aggregate is required in the construction of the GBS. Previous sampling in the area by Bragg (1991) resulted in locating suitable bedrock for high-quality aggregate potential close to Great Mosquito Cove. The present study was undertaken to geologically map the area in order to define the extent and thicknesses of suitable bedrock units and to test representative samples from selected rock types within these units to verify their quality.

In order for a particular rock type to be considered as a high-quality aggregate for concrete use it must pass rigorous physical and chemical testing procedures. Physical testing such as the Los Angeles Abrasion (ASTM C131-89), Magnesium Sulphate Soundness (ASTM C88-83) and specific gravity (ASTM C127-88), and chemical testing such as alkali-aggregate reactivity (Oberhoster and Davis, 1986) were performed on selected samples. Petrographic analyses were done on all samples from the field area.

FIELD WORK

Field work consisted of detailed mapping of the geological units found in the area. Each site investigation included, rock identification, representative sampling, determination of overburden type and thickness, the description of geological features, and determination of petrographic number, following the methods outlined by Bragg (1989).

GEOLOGY

The map area is underlain by a succession of late Precambrian volcanic and sedimentary rocks that are part of the Musgravetown Group (Hayes, 1948), which consists of a lowest formation of mainly volcanic rocks called the Bull Arm Formation (McCartney, 1967) conformably overlain by a sequence of sedimentary rocks.

Within the map area (Figure 1), the Bull Arm Formation (Units 1 to 3) consists of interbedded mafic and felsic volcanic rocks. The mafic rocks (Unit 1) consist of dark-grey, green and purple basaltic flows, tuffs and agglomerates. In places, the rocks are vesicular and amygdaloidal. Amygdules are generally filled with calcite, epidote, chlorite or quartz. Epidote also occurs as veins and in the matrix of the basaltic tuff. The dacitic rocks (Unit 2) consist of light-green crystal-lithic tuffs and ash flows. The felsic rocks (Unit 3) consist of buff, purple and red rhyolitic crystal-lithic tuff, breccia and ash flows. Flow banding, autaxitic structures and zones of lithophyses are common. Crystals in the acidic tuffs are generally K-feldspar, albite and quartz. The volcanic rocks form belts of varying composition that strike in a northwesterly direction and dip moderately (20 to 35°) to the

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[1] Mineral Deposits Section

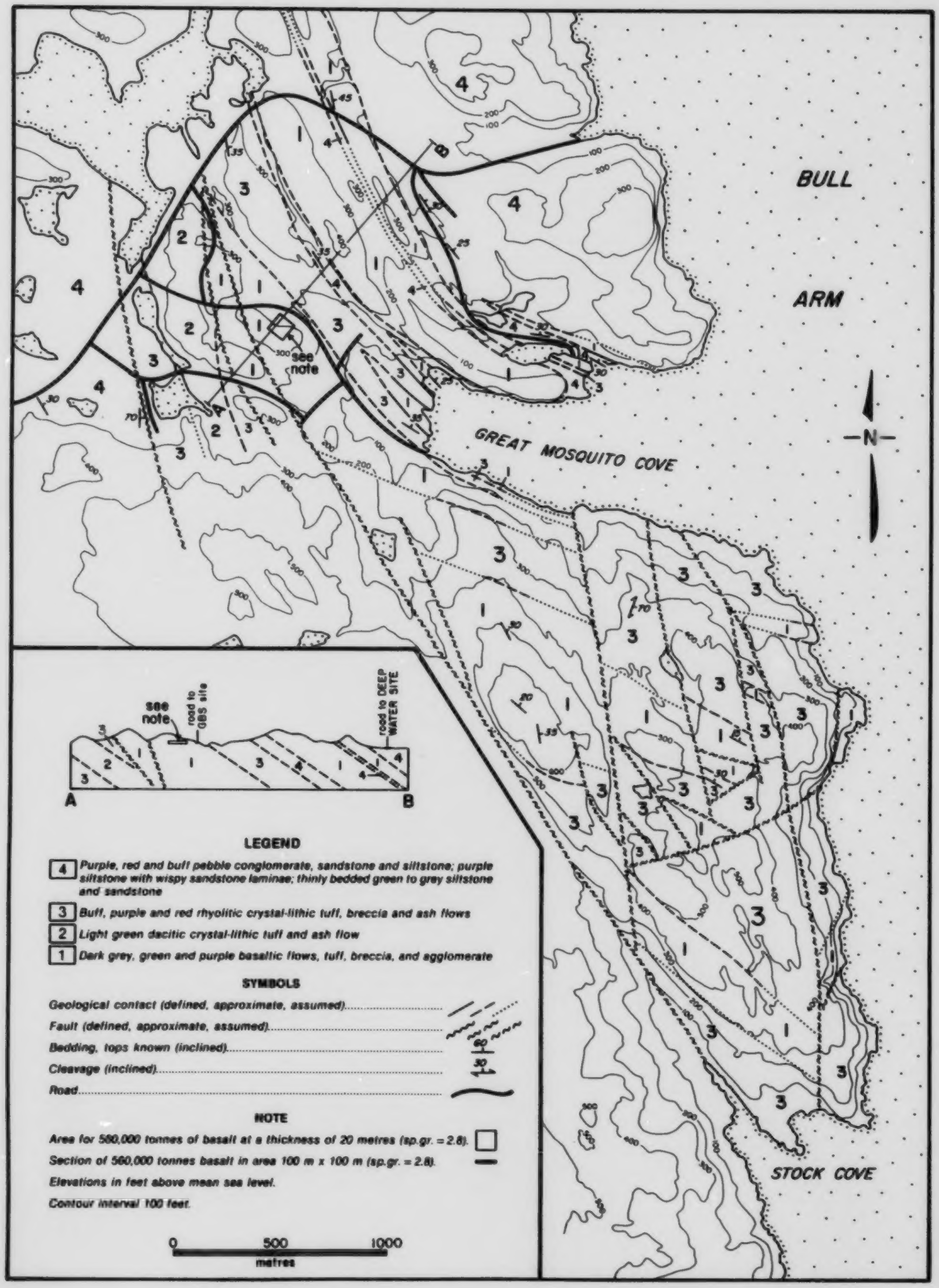
BULL
ARM
GREAT MOSQUITO COVE
STOCK COVE
N
see note
see note
road to GBS site
road to DEEP WATER SITE
A
B
LEGEND
4 Purple, red and buff pebble conglomerate, sandstone and siltstone; purple siltstone with wispy sandstone laminae; thinly bedded green to grey siltstone and sandstone
3 Buff, purple and red rhyolitic crystal-lithic tuff, breccia and ash flows
2 Light green dacitic crystal-lithic tuff and ash flow
1 Dark grey, green and purple basaltic flows, tuff, breccia, and agglomerate
SYMBOLS
Geological contact (defined, approximate, assumed)
Fault (defined, approximate, assumed)
Bedding, tops known (inclined)
60
Cleavage (inclined)
30
Road
NOTE
Area for 560,000 tonnes of basalt at a thickness of 20 metres (sp.gr. = 2.8).
Section of 560,000 tonnes basalt in area 100 m x 100 m (sp.gr. = 2.8).
Elevations in feet above mean sea level.
Contour interval 100 feet.
0
500
1000
metres

Figure 1. *Geology of the Great Mosquito Cove area.*

Table 1. Petrographic number ranges of different rock units found in the Bull Arm–Mosquito Cove area

Group/ Formation	Number of samples	Petro. number range	Petro. number < 150	Petro. number > 150	Average petro. number
Musgravetown Gp. (Unit 4)	16	110-325	10	6	185
Bull Arm Fr.					
(Unit 1)	25	110-140	25	0	130
(Unit 2)	3	110	3	0	110
(Unit 3)	14	110-185	12	2	140

Table 2. Results of Magnesium Sulphate Soundness testing on representative samples from the Mosquito Cove area

Unit #	3	1	1	4	3	1
Sample #	90-001	90-006	90-010	90-013	90-025	91-001
Rock Type	tuff	basalt	basalt	sandstone	rhyolite	basalt
% Loss Fine	2.453	1.63	1.764	3.213	2.63	1.654
% Loss Coarse	1.321	.793	.835	1.841	1.231	.813

northeast. The thicknesses of individual units vary from a few metres to 300 m. North- to northwest-trending faults crosscut the area and offset some of the units.

Sedimentary rocks (Unit 4) of the Musgravetown Group gradationally overlie the volcanic rocks in the northeast part of the map area and are in fault contact to the west. They consist of red to purple, pebble to cobble conglomerates up to 70 m thick are interbedded with and overlie the upper sections of the volcanic rocks. The conglomerates grade upward through red and buff, well-bedded sandstone and siltstone, to very finely bedded grey and green siltstone and sandstone.

LABORATORY INVESTIGATION

Petrographic examination (ASTM C295-85) was carried out on all hand specimens and thin sections. Petrographic examination involves identification of the different rock types and the determining the quality of the rock sample by identification of microstructures (fractures in the crystals) and degree of alteration from weathering and/or metamorphism. Also, a more reliable petrographic number (P.N.) (Bragg, 1989) assessment can be obtained from a petrographic analysis. Table 1 shows petrographic number ranges of different rock units found in the Great Mosquito Cove area as obtained from petrographic examination.

Magnesium Sulphate Soundness Test (ASTM C88-83), which is used to indicate the durability of an aggregate in relation to weathering agents, was performed on selected samples from the area (Table 2).

Los Angles Abrasion Test (ASTM C131-89) for small-size, coarse aggregate was also conducted on the selected samples from the area Table 3. This test is used to measure the durability of an aggregate in relation to wear, impact and abrasion from other agents.

Alkali-Aggregate Reactivity (Oberholser and Davies 1986), which is a 14-day accelerated mortar-bar test was performed on representative samples of the major rock units (acidic crystal-lithic tuff, basalt, basaltic tuff, sandstone and flow-banded rhyolite) and of selected samples from the field area Table 4. This test was chosen because it obtains results quickly (14 days) and it is the most severe test known; if a sample passes this test, no further testing may be necessary. Samples failing the test will require the longer (1 year) Mortar Bar (ASTM C-227) or Concrete Prism (CSA A23.2-14A) tests.

Density and absorption (ASTM C127-88) was performed on selected samples from the basalt unit in the Bull Arm Formation Table 5.

RESULTS AND DISCUSSION

A total of 65 sites were visited and 58 samples collected. A petrographic examination was performed on all hand specimens and thin-sections and an average petrographic number for total samples from each unit is shown in Table

Table 3. Results of Los Angles Abrasion testing on representative samples from the Great Mosquito Cove area

Unit #	3	1	1	4	3	1
Sample #	90-001	90-006	90-010	90-013	90-025	91-001
Rock Type	tuff	basalt	basalt	sandstone	rhyolite	basalt
% Loss L.A. Abrasion	13.4	10.12	11.2	17.8	11.4	11.1

Table 4. Alkali–Aggregate Reactivity of difference rock types in the Bull Arm–Mosquito Cove area

Site #	Rock Type	% Expansion	Comments
Representative Samples			
90-001	Crystal lithic tuff	0.365	Reactive
90-008	Basalt	0.020	Non-reactive
90-010	Basalt	0.029	Non-reactive
90-013	Sandstone	0.404	Reactive
90-025	Flow-banded rhyolite	0.078	Non-reactive
Selected Samples			
91-001	Basalt	0.053	Non-reactive
91-002	Basalt	0.037	Non-reactive
91-003	Crystal lithic tuff	0.254	Reactive
91-004	Basalt	0.015	Non-reactive
91-005	Lithic tuff	0.286	Reactive
91-006	Basalt	0.018	Non-reactive

1. A petrographic number is an initial quality assessment of aggregate or rock sample. Each rock type is assigned a petrographic factor of either 1(best), 3(fair), 6(poor) and 10(deleterious) (CSA A23.2.30, 1973). A petrographic number limit of 135 and 150 is recommended for concrete and asphalt mix, respectively, in Newfoundland. Of the 58 samples collected, 50 samples are considered to be of high quality (P.N. <150); these are mainly from the mafic (Unit 1, 25 samples) and acidic (Unit 3, 12 samples) and intermediate (Unit 2, 3 samples) volcanic rocks from the Bull Arm Formation and the remaining 10 samples are from the Musgravetown Group (Unit 4); 5 samples are considered to be of marginal quality (P.N. range from 150 to 300); 3 samples from Unit 4 (Musgravetown Group) and 2 samples from Unit 3 (Bull Arm Formation). The remaining 3 samples considered to be of poor quality (P.N. >300), and are from Unit 4 of the Musgravetown Group.

Magnesium Sulphate Soundness testing was performed on 6 representative samples from the area (Table 2). A limit of 10 to 15 percent loss is considered maximum for a high-quality aggregate. Table 2 shows that all representative samples from the area passed well below this limit.

Los Angles Abrasion testing was also performed on the 6 representative samples from the Great Mosquito Cove area (Table 3). A limit of 35 percent loss is considered maximum for high-quality aggregate for the above test. Table 3 shows that all representative samples from the area passed well below the limit.

Alkali-Aggregate Reactivity was performed on 5 representative samples and 6 selected samples from the Great Mosquito Cove area (Table 4). An expansion > 0.15 percent at 14 days is considered deleterious and thus reactive for this test. Table 4 shows that the basalt (Unit 1) from the Bull Arm Formation is non-reactive and the flow-banded rhyolite (Unit 3) is non-reactive. The acidic tuffs (Unit 3) from the Bull Arm formation are all potentially reactive, as is the sandstone (Unit 4) from the Musgravetown Group.

Density and Absorption testing was preformed on selected samples from Unit 1 of the Bull Arm Formation in the Mosquito Cove area (Table 5). Aggregates may be classified as low density (< 2.5), normal density (2.5 to 2.9) and high density (>3.0) (Dolar-Mantuani, 1983). The selected samples from the area range from 2.7 to 2.9 having an average of 2.8, and are thus considered normal-density aggregate. The absorption limit varies from 1 to 5 percent depending on use of the aggregate. The selected samples gave results below 1 percent and are thus considered low-absorption aggregates and suitable for all aggregate uses.

CONCLUSIONS

Results from the physical and chemical testing indicate that the basaltic rock types in the area are of high quality and are excellent for use in concrete. This sequence of basaltic rocks is shown as Unit 1 in Figure 1 and can be found in a number of easily accessible areas in sufficient quantities for immediate use. The quality and volume of material in the area suggests that the site may offer potential for aggregate export (Bragg, 1990). The remaining rock types in the area passed the physical testing and are considered excellent sources for road-construction aggregates.

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Table 5. Results of Density and Absorption testing of selected samples from the Mosquito Cove area

Unit #	1	1	1	1	1	1
Sample Number	1	2	3	4	5	6
Rock Type	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
Bulk Relative Density	2.814	2.914	2.790	2.790	2.834	2.825
Absorption	0.190	0.303	0.337	0.195	0.188	0.129
Average Bulk Relative Density	= 2.827					
Avergate Absortion = 0.223						

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QUATERNARY GEOLOGICAL MAPPING, SOUTHWESTERN AVALON PENINSULA

N.R. Catto
Department of Geography
Memorial University of Newfoundland, St. John's, A1B 3X5

ABSTRACT

Mapping of the southwestern Avalon Peninsula, at 1:50 000 scale, has resulted in the identification and delineation of nine major types of geological units. Glacigenic deposits are the most common surficial materials. These diamictons form veneers and blankets over bedrock, and also compose hummocks, flutings, drumlinoid features, and Rögen moraines. The distribution and orientations of these glacigenic landforms indicate that glacial ice accumulated over the northern St. Mary's Bay area and flowed radially to the southwest, west, northwest, and northeast. Separate glaciers occupied the Isthmus of Avalon. The northwesternmost part of the Dildo map area was covered by ice originating from the northwestern side of Placentia Bay.

Exposed and veneered bedrock represent the second most commonly exposed surficial unit. In coastal areas, and in the southern parts of the peninsula, bedrock is the most common material on the surface. Organic deposits are also common throughout the southwestern Avalon Peninsula. Fibric, mesic, and humic sediments are associated with fens and bogs developed over bedrock and glacigenic sediments. Exploitation of peat is presently occurring in several areas. Colluvial, fluvial, marine, glaciofluvial, aeolian, and lacustrine sediments and landforms occupy lesser areas. Glaciofluvial deposits are currently being exploited for aggregate.

INTRODUCTION

Between 1989 and 1991, a series of investigations of the Quaternary geology of the southwestern Avalon Peninsula were undertaken, building on the pioneering work of Henderson (1972). The areas of Dildo (NTS 1N/12), St. Mary's (NTS 1K/13), St. Brides (NTS 1L/16), St. Shotts (NTS 1K/12), Argentia (NTS 1N/5), Placentia (NTS 1N/4), and Ship Cove (NTS 1M/1) have been mapped in detail (Figure 1). This work involved mapping from aerial photographs, followed by detailed stratigraphical and sedimentological field investigations and laboratory analyses of the sediment texture and mineralogical composition. The detailed surficial geology maps produced from this work are designed to be of practical use to industry, government, and the general public.

This paper is intended to accompany the maps, and discusses the composition, material properties, and genesis of the major types of Quaternary geological units identified throughout the seven map areas.

Intensive investigations of the Quaternary sedimentology and stratigraphic framework of the southwestern Avalon region have been actively ongoing since August 1989. Recent data obtained during the 1991 field and laboratory investigations, which resulted in the identification of the ice-flow directions and the extent of interaction between several ice masses centred over St. Mary's Bay, the Avalon Isthmus, and ice originating from glaciers centred in south-central Newfoundland, will be discussed extensively in a forthcoming publication by this author.

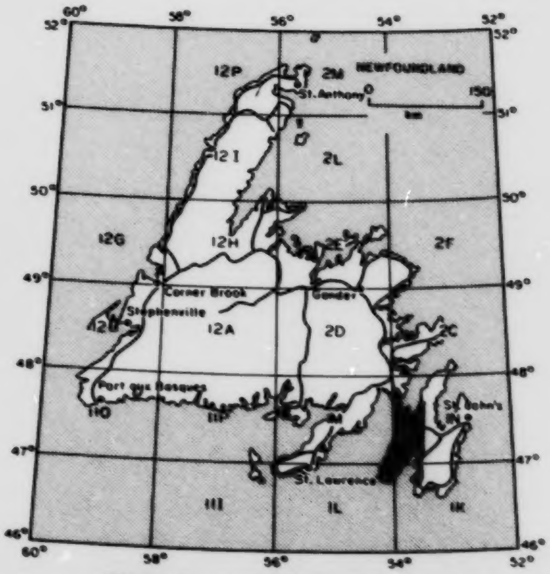

Figure 1. *Location of study area.*

DESCRIPTION AND ANALYSIS OF QUATERNARY GEOLOGICAL UNITS

The terminology and symbols used for designation of the map units uses the system developed for Quaternary geological mapping by the Newfoundland Department of

Mines and Energy. Map units are classified in terms of dominant genetic process and surface expression.

Throughout the southwestern Avalon Peninsula, examples of complex sedimentary successions were encountered along incised valleys adjacent to the coastline, or in areas where mass-movement activity and alluvial reworking was prevalent. In these areas, the symbols are unavoidably complex, and the areas represented by each map unit relatively small.

In locations where two or more sediment types are exposed within the boundaries of a single map unit, a composite unit symbol is used, with the dominant unit listed first. In areas where a unit is stratigraphically beneath a veneer or blanket of another, but influences the topographic expression or tonal qualities visible on aerial photographs, a compound symbol is used with the underlying unit symbol in the denominator.

The number of map units represented by distinct terrain classification symbols precludes discussion of each individual unit. The discussion, which follows, is therefore based upon the dominant genetic processes that each unit type represents. The unit types are discussed in order of diminishing prominence and areal extent.

1) Glacigenic Units (T)*

Sediments mapped and interpreted as glacigenic units are the most common surficial material in the region. These units are extensively exposed throughout. In coastal areas, glacigenic sediments commonly are veneered or blanketed by marine, fluvial, colluvial, or aeolian deposits.

Sediments mapped and interpreted as glacigenic are diamictons, containing significant quantities of sandy silt, sand, and larger clasts. These diamictons were formed by several genetic processes, including: 1) direct deposition from ice during glacial advance or recession (true glacigenic units or 'tills'); 2) deposition by sediment gravity flow beneath or on the surface of glacial ice; 3) deposition from the ice margin into marine waters by sediment gravity flow or other mass-movement processes; and 4) initial deposition by one of these three processes, followed by substantial modification in a marine environment.

These qualifications allow the glacigenic units to be treated for geological mapping in terms of their geomorphic expression and texture, without specifying the precise mode of genesis. Although diamicton units that have undergone any substantial modification after deposition from glacial ice cannot be classified as tills, and are technically not glacigenic deposits, the units throughout the southwestern Avalon region have broadly similar physical properties. Therefore, for practicality and ease of general discussion, all of the diamicton units which predate marine, fluvial, or aeolian deposition were mapped as glacigenic units, and will be treated collectively in this paper. Diamictons deposited by postglacial colluviation are treated separately, as colluvial deposits.

Depositional analysis of the diamicton sediments in the southwestern Avalon Peninsula is a complex process requiring intensive examination of numerous individual outcrops. Intensive sedimentological analyses are currently being undertaken.

Texturally, the diamictons vary in matrix composition from sandy silt to fine granules. The proportion of pebbles and larger clasts commonly varies between 35 and 65 percent. Throughout the region, the pebbles and cobbles are derived from locally outcropping Upper Proterozoic and Cambrian rock units.

The internal structures within the diamictons show a wide range of styles. Although most of the diamictons are generally texturally homogeneous, without distinctive lenses or stratification, some beds contain sand and silt lenses. The lenses are commonly thin (less than 10 cm thickness) and laterally discontinuous (less than 50 cm maximum dimension). The majority are internally structureless, although rare examples of structures indicative of flowing water (such as crudely defined cross-stratification and graded bedding) are rarely present.

At several exposures, the alignment of the large pebbles and cobbles was measured, in order to ascertain the direction of sediment transport and to assist in interpretation of the genesis of the diamictons. Results from these clast-fabric analyses were variable for the region's diamictons, with vector orientation strengths varying from weak (principal eigenvalue = 0.4) to well-oriented (principal eigenvalue = 0.8). These variations indicate that the sediments mapped here as 'glacigenic' or 'till' represent many styles and processes of deposition.

Veneers (symbol 'v') and blankets (symbol 'b') are the most common surface expressions involving morainic material, especially in interior areas. The topography of these sediments reflects that of the underlying bedrock. Slopes vary from gentle to moderate. Most of the deposits are dominated by coarse pebbles and cobbles, and have undergone some measure of post-depositional weathering.

Many morainic units throughout the region are marked by hummocky (symbol 'h') or eroded and dissected hummocky (symbol 'he') surfaces. The hummocks are generally poorly defined, and range from 2 to 10 m in height and from 5 to 30 m in width. Smaller hummocks are more common. Areas of hummocky topography are marked by gentle to moderately steep slopes.

Flutings, or oriented lineations of diamicton, are represented by the symbol 'l'. Flutings are present in several parts of the region, but are particularly prominent in the northern parts of the Argentia map area, and in the Dildo map area. The largest flutings have maximum heights of 8 m, widths exceeding 20 m, and lengths in excess of 300 m. Slopes are moderate to steep.

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* Symbols used in this paper refer to a forthcoming map publication by the author.

Areas of drumlinoid topography (symbol 'd') are also present in the southwestern Avalon, but are less prominent than the fluting areas. Drumlinoid features are relatively poorly developed, and many have bedrock cores. Maximum heights and widths reach 7 m and 15 m respectively. Slopes are moderate to steep. The orientations of the flutings and drumlins throughout the region indicate formation beneath a glacier that formed over the northern part of St. Mary's Bay and flowed toward the west, southwest and north. This pattern of flow conforms with those determined from the striation orientations, from clast-fabric studies, and from consideration of the lithology of the clast assemblages in the diamictons. Areas of ridged topography (symbol 'r') are also present throughout the region, but are particularly prominent in the eastern part, in the vicinity of Whitborne. The geomorphic form, marked by crescentic ridges oriented with the horns of the crescents generally aligned parallel to the local ice-flow direction, indicates that these features are subglacially formed Rögen moraines. The Rögen moraines have maximum heights of 12 m, and maximum lengths of 100 m. Slopes are moderate to steep.

The precise mechanism of formation of Rögen moraines is a matter of considerable controversy. Further research is currently being undertaken to attempt to resolve the problem of Rögen moraine origin, and to clarify the relationship between their orientation and striation and diamicton clast-fabric measurements.

2) Exposed and Veneered Bedrock ('R')

Exposed and veneered bedrock represent the second most common surface material in the southwestern Avalon Peninsula. The bedrock units throughout the region have been mapped previously by King (1988).

In many coastal areas, and in the southern parts of the region (St. Bride's and St. Shott's map areas), exposed bedrock dominates the surface. In these locations, the bedrock commonly is capped with a thin veneer of broken clasts derived from frost weathering (symbol 'Rw'). The presence of distally derived clasts within the frost-weathered mantles, however, indicates that the areas have been glaciated during the Quaternary.

3) Organic Units (O)

Organic units are defined as those composed largely of organic materials resulting from the accumulation of plant material, containing at least 30 percent organic material by weight (Expert Committee on Soil Survey, 1987). They are associated with fens and bogs developed above relatively impermeable bedrock and glacigenic materials. The surfaces of the units are generally level, forming veneers ('v') and blankets ('b').

The organic deposits are classified according to the system of the Expert Committee on Soil Survey (1987). Fibric deposits are the most common surface forms. The majority of the fens are dominated by *Picea mariana* (black spruce) and *Larix laricina* (tamarack), and often contain standing water. Peat deposits from these fens are dominated by weakly decomposed (1 to 3 on the Von Post scale, as used by the Expert Committee) spruce, tamarack, and sedge fragments. Peat deposits associated with these fens are thin (generally less than 2 m), and are not suitable for agricultural or heating purposes.

Mesic deposits, associated with active and formerly developed bogs, display an intermediate degree of decomposition (4 to 6 on the Von Post scale). These units are composed of decomposed arboreal fragments. Active exploitation of these deposits, along with scattered humic units (those with Von Post decomposition values of 7 or greater), is occurring at several locations in the southwestern Avalon Peninsula, notably in the St. Shotts area.

4) Colluvial Units (C)

Units mapped as colluvial are predominantly diamictons. Surface washing, and to a lesser extent sorting during colluviation, has resulted in a concentration of coarser clasts on the surfaces of many colluvial units. The colluvial units are thus generally coarser than the glacigenic diamictons.

The colluvial sediment is derived from a number of sources, but most deposits incorporate glacigenic material. Marine, fluvial, and previously colluviated sediment is also commonly involved in mass movements in the short, steep valleys that mark the Placentia Bay coastal zone. The geomorphic expression of these units is somewhat variable, but most are roughly undulating or hummocky blankets and veneers, having steep or moderately steep slopes. Eroded surfaces are common, and active colluviation occurs in many localities.

5) Fluvial Units (F)

Fluvial units are produced in the narrow, steep valleys of braided streams throughout the region. Geomorphically, the deposits commonly form veneers or blankets over glacigenic materials or bedrock. Fluvial plains (symbol 'p') are present in the bottoms of the largest valleys, in areas adjacent to the coast.

Texturally, the fluvial deposits vary from medium sand to grain-supported pebble and cobble gravel. Fine gravel units are the most common types present. Generally, the deposits are moderately to well sorted. All of the mappable fluvial units were formed postglacially.

6) Marine Units (M)

Marine units represent both deposits formed during higher sea-level stands (terraces, symbol 't'; and eroded terraces, symbol 'te'), and deposits currently forming at sea level (beach ridges, symbol 'r'). All of these deposits are dominated by coarse sand and gravel, and fine marine sediments are not present in successions associated with high sea-level episodes.

The chronology of marine events in the region is still unclear, and investigations are ongoing. Evidence collected suggest that sea levels were higher than present during the latest stages of deglaciation and during the early Holocene, 10,000 to 8,000 years BP. The highest marine deposits and terraces indicate sea levels approximately 15 to 20 m above present. Subsequently, sea level dropped below the present surface. Currently, sea level is slowly rising around the southwestern Avalon, as indicated by submerged terrestrial peat deposits at several localities.

7) Glaciofluvial Units (G)

Glaciofluvial landforms include ridged esker deposits (symbol 'r'), esker and fan-delta blankets and aprons overlying glacigenic materials and bedrock (symbols 'b' and 'a'), and eroded gravel and sand units (symbol 'e'). These deposits, composed of sand and gravel, represent the main sources of aggregate in the region. The largest glaciofluvial deposit, the esker fan delta at Point Verde, is currently being exploited for aggregate, and other deposits between Placentia and Colinet have been exploited in previous years.

The glaciofluvial deposits represent sedimentation by streams associated with glaciers in subglacial, supraglacial, and ice-marginal environments. Analysis of the environments of sedimentation of these deposits is currently in progress.

8) Aeolian Units (E)

Aeolian units are uncommon in the southwestern Avalon Peninsula. Dunal deposits are largely confined to Lance Cove. The dunes are small, poorly developed domal features, formed by onshore winds. Sediments associated with the dunes are well- to moderately-sorted sand, generally in structureless beds. Extensive modification by lesser velocity winds has resulted in the development of blowouts over the dune surfaces. Aeolian erosion continues today in unvegetated or cleared areas.

Thin veneers of aeolian silt and silty sand loess occur as caps over marine, glacigenic, and glaciofluvial sediments. These cliff-top loess deposits are thickest at the cliff edges, decreasing exponentially in thickness with distance from the escarpment. Generally, they occur only within 30 m of cliff edges.

9) Lacustrine (L)

Units mapped as lacustrine were deposited in shallow lakes that do not owe their existence to blockage of the regional drainage by glaciation. The lacustrine units, composed of silt and silty clay, are of very limited areal extent. The surface expression commonly is a small, featureless plain, bordering an existing pond.

SUMMARY

Mapping of the southwestern Avalon Peninsula at 1:50 000 scale has resulted in the identification and delineation of nine major types of geological units. Glacigenic deposits are the most common surficial materials. These diamictons form veneers and blankets over bedrock, and also compose hummocks, flutings, drumlinoid features, and Rõgen moraines. The distribution and orientations of these glacigenic landforms indicate that glacial ice accumulated over the northern St. Mary's Bay area and flowed radially to the southwest, west, northwest, and northeast. Separate glaciers occupied the Isthmus of Avalon. The northwesternmost part of the Dildo map area was covered by ice originating from the northwestern side of Placentia Bay.

Exposed and veneered bedrock represents the second most commonly exposed surficial unit. In coastal areas, and in the southern parts of the peninsula, bedrock is the most common material on the surface.

Organic deposits are also common throughout the southwestern Avalon Peninsula. Fibric, mesic, and humic sediments are associated with fens and bogs developed over bedrock and glacigenic sediments. Exploitation of peat is presently occurring in several areas. Colluvial, fluvial, marine, glaciofluvial, aeolian, and lacustrine sediments and landforms occupy lesser areas. Glaciofluvial deposits are currently being exploited for aggregate.

Ongoing research in the southwestern Avalon involves assessment of ice-flow directions and relative chronology, analysis of the environments of formation of glacigenic sediments, investigation of the history of sea-level changes, and aspects of formation and development of marine, glaciofluvial, and aeolian deposits.

ACKNOWLEDGMENTS

This research was supported by the Natural Sciences and Engineering Research Council of Canada. I am indebted to Dr. Dave Liverman, Martin Batterson, and all my colleagues of the Newfoundland Department of Mines and Energy, for extensive consultations during all stages of the research. Capable field assistance was supplied by G. Thistle.

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SURFICIAL GEOLOGY OF THE SCHEFFERVILLE AREA (LABRADOR PARTS OF NTS 23J/10 AND 23J/15)

D. Liverman and H. Vatcher
Terrain Sciences Section

ABSTRACT

Surficial geology mapping was conducted in the Schefferville area (the Labrador parts of map sheet NTS 23J/10 and 23J/15). Detailed ice-flow mapping in this study provides support for the previous models of regional ice-flow history. The results suggest that such models may be applied on a local scale, although some local topographic variation from regional patterns is apparent in areas of high relief. Where well-oriented clast fabrics are found, they suggest northwest–southeast flow indicating that a later northeast flow had little effect on deposited sediment.

The landscape in the study area is dominantly one of erosion, rather than deposition, and this will result in problems when applying conventional drift-exploration methods. Over much of the area, the effects of meltwater erosion and deposition dominate, with glacial processes having little effect on the surface sediment cover. Thus, both glacial and glaciofluvial models of dispersal have to be considered.

Thick sections exposed in mine pits are interpreted as showing the effects of the last glacial event to affect the area, rather than a long glacial history. Surface diamictons show no consistent colour, but reflect the erodibility and colour of local bedrock.

INTRODUCTION

The Schefferville area straddles the border between western Labrador and northeastern Quebec. From 1945 to 1982, it was the site of intensive exploration, development and mining of iron ore. Recent research by the Department of Mines and Energy has identified the Labrador Trough as a likely area for other mineral resources including PGE, and massive sulphides (Swinden *et al.*, 1991). Studies by the Geological Survey of Canada have identified this region as an area of complex ice-flow, and possibly possessing an interesting Quaternary stratigraphic record (Klassen and Thompson, 1987; *in press*). This, combined with the comparative cost advantages of work in a relatively developed area, led to selection of the Newfoundland part of the 23J/10 and 23J/15 NTS map sheets for study.

OBJECTIVES

The objectives of the project are:

1) map the surficial sediment cover of the area, using aerial photographs and field checking; and to identify the grain size, thickness and genesis of the sediment;
2) map ice-flow indicators (striations, landforms, and clast fabrics) and investigate clast provenance in order to examine the ice-flow history of the area;
3) examine and map the geomorphology of the area;
4) integrate the information obtained to interpret a Quaternary geological history for the area; and
5) use the geological history in developing a model for sediment dispersal that would assist mineral-exploration companies using drift-exploration techniques.

LOCATION AND ACCESS

The town of Schefferville is located 150 km north of Labrador City in the interior of the Labrador–Ungava Peninsula (Figures 1 and 2). It is serviced by the Q.N.S.& L. rail service, and a modern airport, but has no external road connections. Access in the northern part of the map area is good via the network of mining and exploration roads built during the period of mining activity. Access to the southern part of the map area is possible by boat or helicopter, whereas the western hills are only accessible by helicopter.

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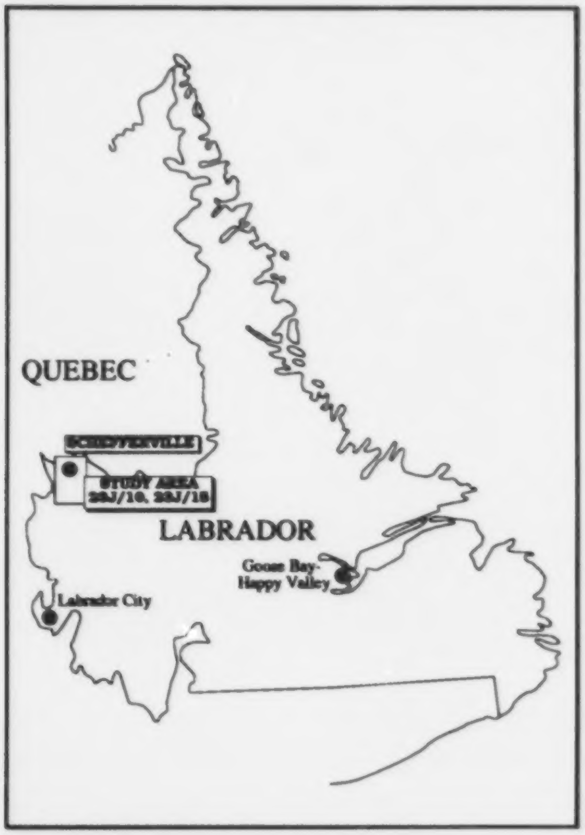


Figure 1. *Location of study area.*

BEDROCK GEOLOGY

The geology of the area was first described by A.P. Low, who also identified the economic potential of the iron-ore deposits of the Labrador Trough (Low, 1896). Further detailed mapping was performed from the 1940's onward by the Iron Ore Company of Canada, and the Geological Survey of Canada. Wardle (1982) compiled existing information on the Labrador Trough geology into a comprehensive map of the area. The bedrock of the southwest part of the study area consists of Archean gneisses and granites of the Superior Province Ashuanapi Complex (Greene, 1974; James and Stephenson, *this volume*). The remainder of the study area is underlain by Lower Proterozoic sedimentary and volcanic supracrustal rocks of the Labrador Trough (Greene, 1974; Wardle, 1979; Wardle and Bailey, 1981). These include iron formation, chert, quartzite, dolomite, sandstone, siltstone and shale of the Sokoman, Denault, Menihek, Fleming, Wishart and Dolly formations (Wardle, 1982).

Economic interest has been mainly in the iron ore exploited for 30 years in the Shefferville area. McConnell (1984) described zinc anomalies found in lake- and stream-geochemistry surveys in the Iron Arm and Howell's River areas, but was unable to locate a source. Swinden *et al.* (1991), on the basis of several prospects and occurrences, suggested that the eastern Labrador Trough has the potential for further massive sulphide deposits.

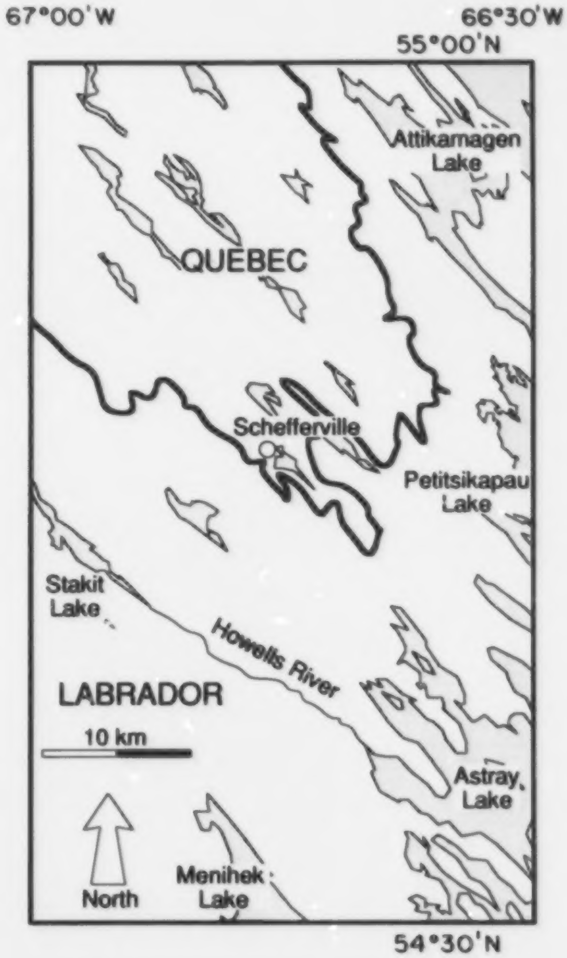


Figure 2. *Main features of the study area.*

PREVIOUS WORK

Low (1896) noted striations at the north end of Dyke Lake indicating southwesterly ice flow. Harrison (1952) recognized three directions of flow in the Schefferville area, a southeastward flow that sculpted the main features of the landscape, followed by a flow to the northwest, and subsequently a flow to the northeast. Henderson (1959) also recognized an early dominant landforming southeasterly flow, with a later shift in flow to the east. He was unable to find evidence for a northeast flow in the Schefferville area, but on the evidence of striae in the Stakit Lake area, suggested that a regional northeast flow preceded a local late southeast flow down the Howell's River valley (Figure 2). Kirby (1960, 1961, 1966) reported till-fabric analyses in the Schefferville area that showed the last ice movement in the area was oriented north-northwest–south-southeast. He also reported striations that are interpreted as showing an early northeast flow followed by a later south-southeast flow, suggesting that the age relationship derived by Henderson (1959) was incorrect. Klassen and Thompson (1987, 1989, *in press*) identified five phases of ice movement in the Schefferville area through mapping of erratics and striation evidence. Phase I consisted of westward flow from the Schefferville area, identified by dispersal of trough erratics onto the Archean highlands to the west of the study area. Phase II consisted of north to north-northeast flow, affecting mostly the area north and west of the present study area. Phase III consisted of flow both northwest and southeast from an ice centre located southeast of Attikamagen Lake, which migrated north of Schefferville. Phase IV consisted of strong eastward flow across the Archean Highlands into the Trough, affecting mainly the area south of Schefferville. Phase V was considered to be minor and short lived, and resulted in northeastward flow in the Schefferville area, which had little effect on outcrop morphology.

Klassen and Thompson (1987) described till stratigraphy in the region. A regionally recognized red to red-brown surface till overlay a grey-brown till, which in turn overlies a dark-grey till. At one site, fluvial gravels were identified separating red and grey-brown tills, possibly indicating preservation of non-glacial sediments. Klassen and Thompson (1987) suggested that the tills are lithologically distinct, and may represent significant changes in the directions of glacial transport.

Henderson (1959), Derbyshire (1960, 1962, 1964) and Ives and Kirby (1964) described the numerous meltwater channels found in the Schefferville area, suggesting that most were formed subglacially rather than at the margin of retreating ice.

FIELD METHODS

The study area was covered using a combination of all-terrain vehicle, boat, helicopter and foot. Striations were measured and mapped from rock outcrops. Existing trenches, pits, and disused mines provided exposure of sediments for section work supplemented by hand excavation of shallow pits. Well-exposed sections were cleared, described, and photographed, and where considered relevant, clast-fabric measurements were made using standard techniques. From diamicton exposures, samples of matrix were collected and analyzed for geochemistry and texture, and 80 to 120 pebbles were collected for clast lithology identification.

A significant departure from field methods used by previous workers was the routine use of a sun compass. Such an instrument is required due to the presence of large volumes of iron-rich rock in the Schefferville area, which disturbs the local magnetic field. Use of the sun compass (see Larochelle, 1963 for description of methods) showed that magnetic compasses were unreliable with unpredictable variations over short distances, with deviations ranging from 0 to 60°, averaging 10 to 20°.

GEOMORPHOLOGY AND SURFICIAL GEOLOGY

The geomorphology and surficial geology of the study area is varied and form distinct zones (Figure 3). The highlands southwest of the Howells River valley have a landscape dominated by ice-sculpted bedrock having numerous rôches moutonées and perched boulders, with the eastern slopes showing many bedrock-controlled meltwater channels. Drainage in this area is disorganized, and reflects bedrock weaknesses. Surficial sediment is generally thin and restricted to valley bottoms, consisting mostly of a veneer of diamicton, with some glaciofluvial deposits, and areas of bog.

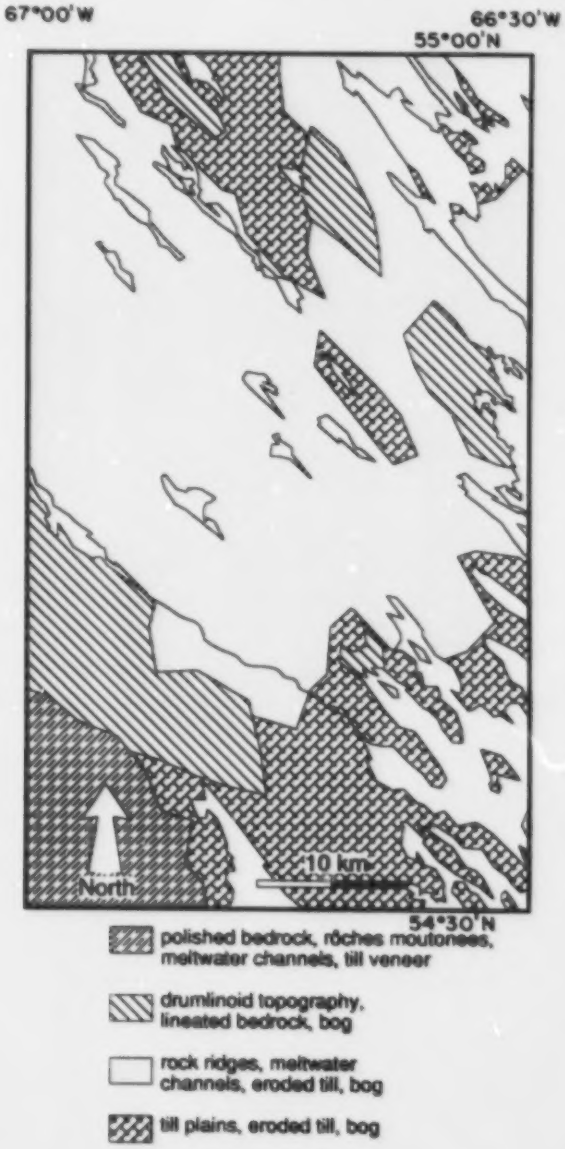

67°00'W
66°30'W
55°00'N
10 km
North
54°30'N
polished bedrock, rôches moutonées, meltwater channels, till veneer
drumlinoid topography, lineated bedrock, bog
rock ridges, meltwater channels, eroded till, bog
till plains, eroded till, bog

Figure 3. *Sketch map showing simplified geomorphology and surficial geology of the study area.*

The Howells River valley and Stakit Lake area form a broad lowland with numerous drumlinoid landforms. Relief is comparatively low (2 to 8 m), and individual features are

200 to 500 m long and 50 to 300 m wide. Limited examination of these features suggests that they are largely composed of bedrock having a thin till cover on the crests, and extensive areas of bog in intervening lows. Their morphology is variable, but generally does not fit the 'classical' egg-shaped drumlin. The long axes of the features are oriented northwest-southeast, parallel to bedrock strike. The genesis of drumlins and related features is controversial. The conventional view is that they form subglacially through the action of ice flow on a till substrate (Smalley and Unwin, 1969; Boulton, 1987; Menzies, 1989; Boyce and Eyles, 1991). This model has been challenged by Shaw (1983), Shaw and Kvill (1984), Sharpe (1987), and Shaw and Ashley (1988) who suggest that drumlins are formed by major subglacial meltwater flows. The morphology of the Schefferville features resembles those described by Shaw and Kvill (1984), and their bedrock composition suggests that these are erosional features. The common regional occurrence of bedrock sculpted by meltwater, and the numerous meltwater channels indicate a very active high-pressure subglacial water system and provide some inconclusive support for drumlin formation by subglacial meltwater, with large-scale water erosion modifying existing bedrock ridges.

Plate 1. *Meltwater channels, southeast of Stakit Lake.*

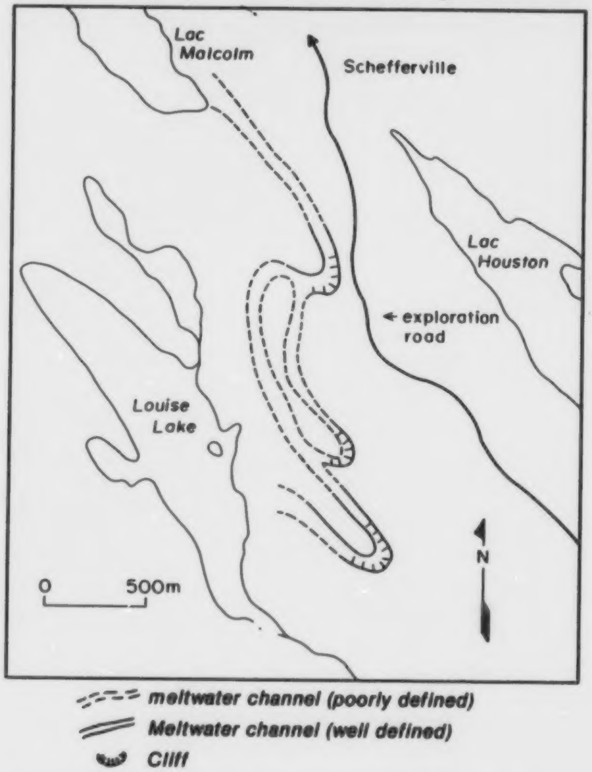


Figure 4. *Sketch map of meltwater channels, Louise Lake area.*

East of the Howells River valley and Stakit Lake, the landscape consists of bedrock ridges, with their sides scarred by meltwater channels (Plate 1). The ridges have some areas of till veneer and thicker areas of till are restricted to valley bottoms or are found in the lee of bedrock highs. Glaciofluvial deposits are rare, and consist mainly of gravel fans at the base of meltwater channels. Meltwater channels are common on valley sides. Two meltwater channels located east of Louise Lake, 10 km southeast of Schefferville, illustrate the control of bedrock structure on morphology. They are found on the western slope of a northwest–southeast-oriented bedrock ridge, and both have an elongate U shape in plan, with the bottom of the U lying to the south (Figure 4, Plate 2). The channels are connected by a poorly defined sinuous channel, with a further U shaped off-shoot. The channels are steep sided, flat bottomed and slope up to the south at 2 to 5°. The height of the outer wall increases toward the apex of the U, reaching an estimated 45 to 60 m high in the northerly example, and 30 to 40 m in the other. The inner walls also increase in height to a maximum of 10 to 20 m at the apex in each case. The width of the channels remains fairly constant at 100 to 200 m, widening at the apex of the U. The area surrounding the channels consists mostly of bare rock and a few scattered erratics. The local bedrock consists of shales and siltstones of the Dolly Formation (Wardle, 1982), and examination of the orientation of bedding planes surrounding the channels shows that each formed within a dipping antiform. The meltwater has preferentially eroded softer shale horizons, and its path was controlled by the structure of surrounding, more resistant siltstones. It is suggested that such landforms must form subglacially, with meltwater under pressure, in order to form channels that run against topographic gradient. The almost 180° change in direction of the channels resulted in particularly high-water pressures and velocities on the outside wall at the apex of the U resulting in maximum erosion at this point. The poorly defined channel linking the features indicate that they formed simultaneously as part of a single subglacial meltwater system, possibly as subglacial meltwater flow was confined against the valley sides. The shape and path of many other meltwater channels in the area can be related to bedrock structure, generally following bedrock strike, and exploiting easily eroded bedrock. Meltwater channels crossing strike do so along faults and other bedrock weaknesses.

Around Astray and Petitsikapau lakes the landscape is subdued, and has a more continuous drift cover. Where fresh polished bedrock outcrops are exposed (notably at lake

Plate 2. *U-shaped meltwater channels southeast of Schefferville; note slope up to apex of U.*

margins), they are well striated, but sinuous grooves, cavettos, and 'rat-tails' are commonly found (Plate 3). These suggest high velocity subglacial water flow during deglaciation, and may be related to meltwater channels and drumlinoid features described above. Sediment cover is variable in thickness, but generally less than 1 to 3 m. Till is common, and is frequently eroded by minor meltwater channels, or forms featureless plains. Glaciofluvial deposits are rare to absent. Areas of bog are extensive at the margins of large lakes, and in any low-lying areas. Howell's River has formed a large delta, 1 to 2 km wide, where it enters Astray Lake, constituting a significant area of modern alluvium.

Plate 3. *'Rat-tails' in sinuous groove, likely formed by subglacial meltwater erosion, north end of Dyke Lake.*

ICE FLOW

Interpretation of ice-flow history is based mainly on mapping of striations and distribution of indicator lithology clasts. Some care has to be exercised when examining outcrop on the margins of major lakes in the area. The effects of wind on lake ice results in major push effects on the lake shore, with boulder ramparts generated on stony beaches, and inhibition of tree growth up to 30 m back from the shoreline (Pyökäri, 1981). Such shore ice may freeze boulders to its base, and striate bedrock as it is thrust over the shoreline. Thus, striae sites at lake edges were examined critically, with the expectation that striations resulting from ice push would be shallower, broader, and less consistent, both locally and regionally, than glacial striae.

Ice flow is complex, with at least three regionally significant flow events clearly recorded (Figure 5). A strong flow to the southeast, parallel to bedrock strike is shown by large-scale oriented landforms (drumlinoid features, crag and tails, and numerous rôches-moutonées). Striations parallel to this flow trend are common, but are often overprinted and crosscut by striae indicating flow to the northeast. In the Attigamagen Lake area, this later flow can be separated into two components, a northeast flow and an east-northeast flow, although the age relationship between them is not clear. Labrador Trough erratics found overlying Ashuanapi Complex bedrock in the southwest of the field area suggest a westward flow although no corroborating striation evidence was found. These observations can be easily accommodated in the flow models of Klassen and Thompson (1987, 1989).

Forty fabric measurements were made in this study, of which only three show strong unimodal fabrics (using the statistical criteria outlined in Liverman *et al.*, 1990). These are all indicative of northwest–southeast flow, supporting the contention of Kirby (1960, 1961, 1966) that this ice flow was the main influence on sediment dispersal.

CLAST FABRIC

Clast fabric is an important indicator of the environment of deposition in diamictons as orientation of clasts is controlled by the processes depositing the diamicton. Diamictons are of particular interest as they are the chosen medium of sampling in most drift-prospecting studies. In lodgment till, strong orientations both parallel and transverse to glacial flow direction are expected (Lawson, 1979, 1982; Dowdeswell and Sharp, 1986). In basal melt-out till, fabrics are also strong, but may be subsequently modified by reworking. Subaerial debris flow results in less well-oriented inconsistent girdle fabrics, sometimes oriented parallel to local slope (Boulton, 1971; Lawson, 1979). When considering the results of drift prospecting, the genesis of sediment being sampled is important, with transport distances varying considerably between basal tills, and diamictons deposited by other processes.

Forty clast fabrics were obtained on diamictons throughout the area, mostly from thick sections exposed in mine pits in the central part of sheet NTS 23J/15. Clast fabrics were analyzed on a Macintosh microcomputer using the Stereo™ program (MacEachren, 1989).

Mean orientation, and strength of alignment were calculated by the eigenvector methods of Mark (1973, 1974) and Woodcock (1977). The S_1 value measures the strength of the mean clast alignment, and can range from 0.33 (weak) to 1.0 (strong). Figure 6 is a plot following the method of

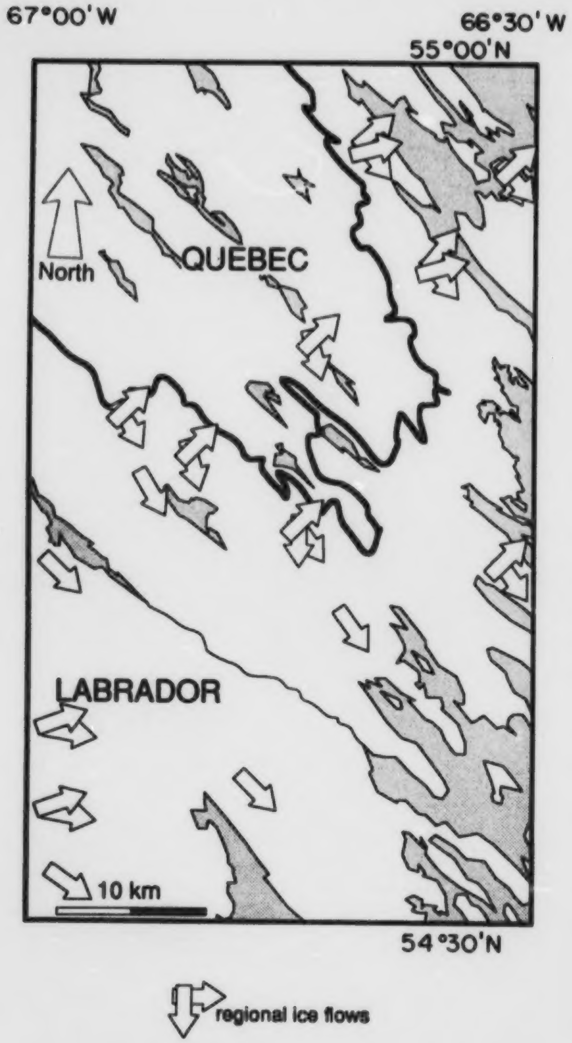


Figure 5. *Ice-flow history, Schefferville area.*

Woodcock (1977) and Rappol (1985). In this diagram, fabrics plotting in the upper left side of the diagram are considered to have a cluster distribution, whereas those in the lower right part have a girdle distribution. The parallel diagonal lines mark increasing values of a strength parameter (k), with weak fabrics plotting closer to the origin. Examination of the diagram shows that fabrics are variable, but weak to moderate girdle fabrics dominate. This indicates that undisturbed basal tills are rare, and the use of clast fabric as an ice-flow indicator requires caution.

Burnt Creek Sections

The thickest and best exposed sections in the study area are found in the Burnt Creek/French mine that straddles the Quebec–Labrador border 4 km north of Schefferville. Up to 30 m of Quaternary sediments are found overlying highly weathered bedrock (Plate 4). Two sections are described in detail here.

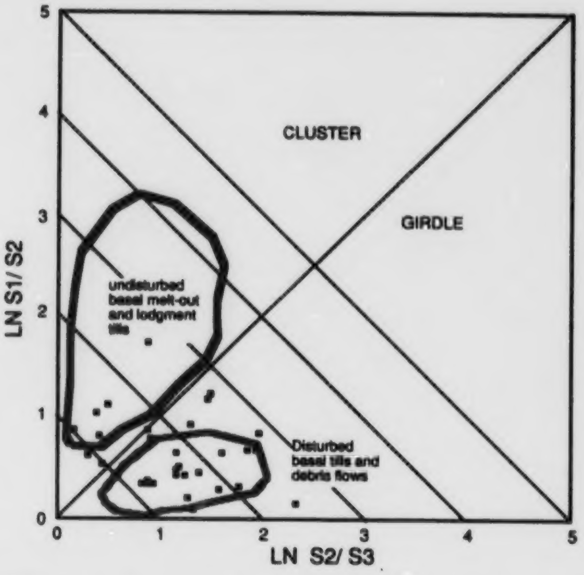
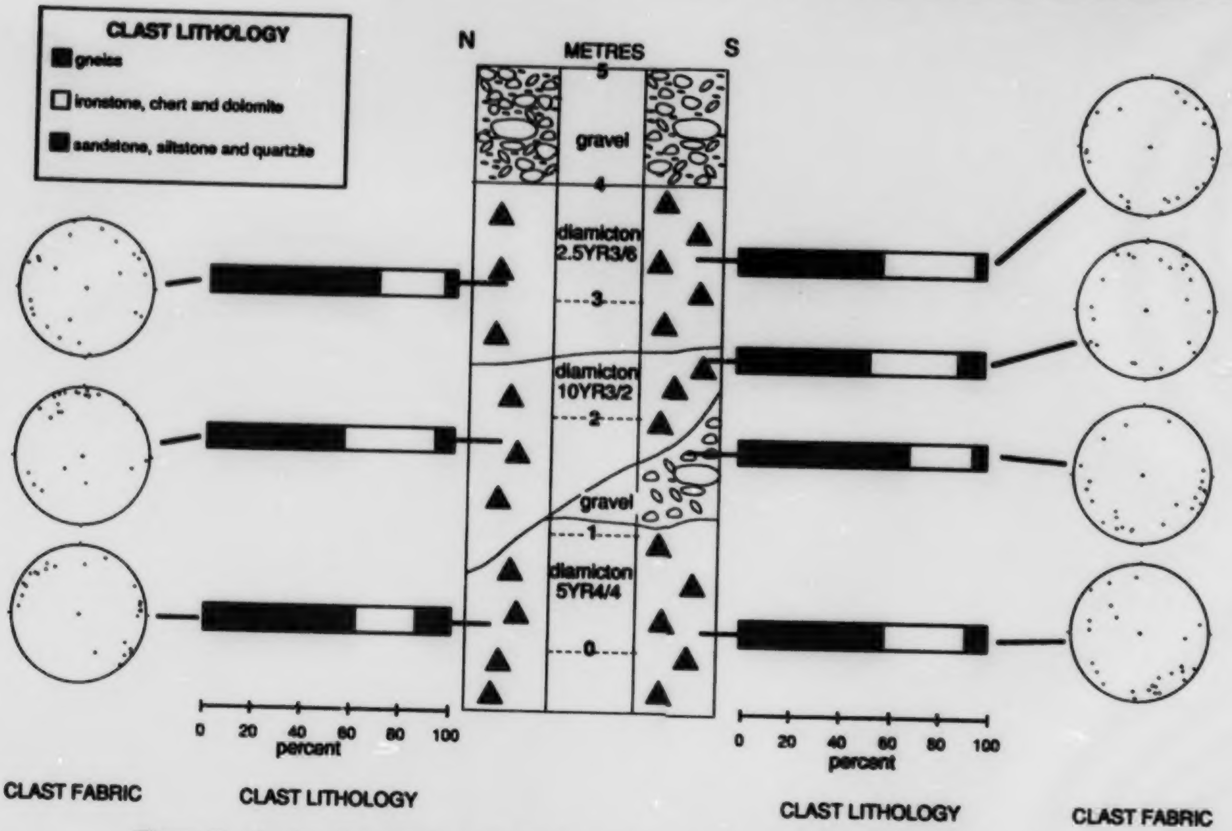


Figure 6. *Diagram of clast-fabric eigenvalues, using the method of Woodcock (1977). The Y axis is the natural logarithm of the ratio between S_1 and S_2, and the X axis is the natural logarithm of the ratio between S_1 and S_3.*

Plate 4. *West wall of Burnt Creek mine, showing up to 30 m of surficial sediment exposed in broad bedrock low.*

exposed, showing three to four diamicton units of differing colour (Figure 7, Plate 5). Two vertical transects, 3 m apart, were made through the section.

The basal unit is a coarse dark reddish-brown diamicton (Munsell colour 5YR 4/4, moist), at least 1.5 m thick. It is massive, unsorted, having a matrix of medium sand to granules, and stone-rich. The clast component consists of 30 percent cobbles to boulders, higher than most diamictons seen in the study area. Two clast fabrics have a general northwest–southeast trend, and show moderately to well-oriented girdle distributions (S_1 values of 0.73 and 0.61).


CLAST LITHOLOGY
▨ gneiss
☐ ironstone, chert and dolomite
■ sandstone, siltstone and quartzite
N
METRES
S
5
gravel
4
diamicton 2.5YR3/6
3
diamicton 10YR3/2
2
gravel
1
diamicton 5YR4/4
0
0 20 40 60 80 100
percent
0 20 40 60 80 100
percent
CLAST FABRIC
CLAST LITHOLOGY
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CLAST FABRIC

Figure 7. *Section diagram, Burnt Creek 1, showing lithology, clast fabrics and clast lithology.*

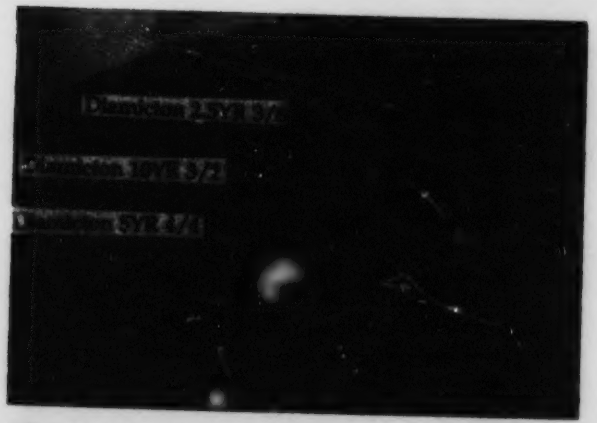
Diamicton 2.5YR 3/6
Diamicton 10YR 3/2
Diamicton 5YR 4/4

Plate 5: *Burnt Creek 1 section. See text for explanation.*

The second unit is only seen on the southern side of the exposure, and consists of a very poorly sorted pebble gravel. It is 1 m thick at this point, but pinches out to the north. It is massive, clast to matrix supported, having cobbles and granules and a coarse sand matrix. Clasts are mostly subrounded. A single clast fabric shows a moderately oriented girdle distribution (S_1=0.54), with an east–west trend.

Overlying a sharp contact that dips at a low angle to the north is a dark greyish-brown diamicton (10YR 3/2, moist) of generally finer matrix texture than the underlying units. The unit is 1 m thick on the north side of the exposure but thins to 30 cm to the south. The matrix is poorly sorted consisting of mostly silt and clay, and clasts compose 50 percent of the unit. Cobbles and boulders are rare, forming less than 5 percent of the coarse clasts. Two clast fabrics trend north–south, and northeast–southwest, and show moderately oriented girdle distributions (S_1=0.55, S_1=0.59 respectively).

The uppermost diamicton unit is 1.2 to 1.5 m thick, and dark red (2.5 YR 3/6, moist). The basal contact is horizontal and sharp. The diamicton is massive, and moderately stony, having very few cobbles or boulders. The matrix is poorly sorted silty clay. Two clast fabrics show weak to moderate girdle distributions (S_1=0.51, 0.55), with southwest–northeast and east–west trends.

The section is capped by 1.5 m of poorly sorted, clast-supported pebble-cobble gravel. Cobbles and boulders form 30 percent of the unit, which has a coarse sand and granule matrix. The basal contact is sharp and horizontal, and is overlain by a lag of cobbles. The gravel shows some rough horizontal bedding, and clasts are mostly subrounded to rounded.

The clast lithology of the section is uniform, with no clear differentiation between units (Figure 7), showing between 4 and 13 percent gneiss derived from the Ashuanapi Complex, 53 to 66 percent clastics, and 25 to 37 percent iron formation, chert and dolomite.

Klassen and Thompson (1987) suggested that the various coloured till sheets in the area may be related to the five-fold ice-flow history they defined by striations and indicator erratics. This section, consisting of multiple diamicton units, offers the potential for such correlation. The last two ice-flow events to affect this area were a strong southeast flow, followed by a late northeast flow. The lowermost unit has consistent, moderate to strong clast orientations, and may be a basal till deposited in conjunction with a southeast ice flow. Fabrics in the overlying diamictons, however, are inconsistent in orientation, and are typical of those obtained from debris flows rather than basal tills. Deposition of these units as a series of debris flows is also suggested by the lateral variability in thicknesses, the similarity in clast lithology throughout the section, and the textural variation. The section lies in a broad bedrock low, and would have acted as a sink for either subglacial or supraglacial gravity flow. This sequence is only found in one location in the pit, despite extensive exposure of Quaternary sediments, suggesting that each diamicton does not represent a regional event. The bedrock exposed in the mine pit is variable in colour, and the colour variations within the section may be due to local variation in incorporation of local material, rather than regional provenance. It is suggested, then, that the diamicton units represent a single basal till, overlain by a series of debris flows. The overlying gravel is coarse and poorly sorted, and was likely deposited in an ice-proximal glaciofluvial environment.

Burnt Creek 2

The thickest exposures of Quaternary sediment in the area lie on the southwest wall of the Burnt Creek/French mine pit. Extensive clearing was required to expose a continuous section, so descriptions were made in a number of gulleys, and incorporated into a composite section (Figure 8).

The thickest sequences lie in a broad bedrock low (Plate 4), with 30 m of surficial sediment overlying an undulating surface of soft and highly weathered bedrock. Bedrock outcrops to the surface on the mine pit wall 350 m north of the described section, but constitutes less than 3 m of the 30-m-high wall at the base of the bedrock low. The lowest surficial unit exposed is a discontinuous bed of well-sorted sand, silt, and clay up to 50 cm thick. It consists mostly of planar-laminated fine sand and silt, draping an uneven bedrock surface. Some indistinct ripple lamination is also present. Rare coarse clasts up to 3 cm diameter are found. This unit is overlain by a diamicton, 1 to 1.2 m thick, very dark greyish-brown to brown (10YR 3/2, 10YR 5/3, moist), containing a single small sand lens. The basal contact is planar and dips to the east at 5 to 10°. The contact undulates, and clasts from the diamicton appear to be impressed into the underlying unit. The diamicton is in turn overlain by a well-sorted sand and silt bed, 10 to 15 cm thick, having irregular, undulating upper and lower contacts. The bed fines upward from medium sand to fine sand and silt. This is overlain by 15 m of diamicton that varies in colour from dark grey to dark brown (10YR 4/1, 10YR 5/2, 10YR 3/2, moist). The upper 10 m of this unit is marked by a network of reddish-brown and brown bands (10YR 4/3, 5YR 4/4, dry). These bands are 3 to 10 cm thick, vary in thickness laterally, mostly lie flat, and are spaced 0.5 to 1 m apart. They surround thin laminae of sand.

Overlying a covered contact is 9 m of yellowish-red diamicton (5YR 4/6, moist), stonier than underlying diamictons and more consolidated. The southern end of the section lies lower than the northern end, and here the lower dark-grey diamicton is directly overlain by 1 to 2 m of coarse gravel. This unit is very poorly sorted, and contains many boulders and cobbles, having a maximum diameter of 1 m. The gravel is clast supported, roughly horizontally bedded, with a coarse sand to granule matrix. The basal contact is sharp and undulating, but has no lag of coarse clasts. The unit thins to a veneer of cobbles and pebbles where it overlies the yellowish-red diamicton.

Ten clast fabrics were obtained from this section. The majority are weak to moderate girdle distributions, having no consistent mean orientation, typical of debris flows (Boulton, 1971; Lawson, 1979). The exception is the uppermost measurement, which shows a strong south to southwest unimodal distribution ($S_1=0.72$), and may represent basal deposition either by lodgement or basal melt-out.

Clast lithology shows little consistent vertical variability, and the pebble samples contain between 58 and 80 percent clastics, 15 and 45 percent iron formation, chert and dolomite, and 1 to 9 percent gneiss. Colour differences between diamictons do not seem to reflect major provenance differences.

It is suggested that the diamicton sequences represent deposition in a large subglacial cavity in the lee of the bedrock high, north of the described section. Large volumes of diamicton and meltwater would be released from the base of the ice at the up-ice edge of the cavity, and would flow to fill the topographic low. The conformable interbedding of sorted sediments with diamicton at the base of the section is interpreted as showing deposition of diamictons by debris flow into standing water. The network of banding seen higher in the section may represent sorting at the margins of debris flows by running water. The colouration is likely formed post-depositionally due to oxidation by percolating groundwater. The cavity would be progressively filled, and then disappear, being overridden by flowing ice. As the ice sheet waned, the upper diamicton was deposited by lodgement or basal melt-out. Subsequent meltwater erosion partially removed the upper diamicton cover, leaving gravels directly overlying the debris-flow sequence.

CONCLUSIONS

The detailed ice-flow mapping in this study provides support for the regional ice-flow history proposed by Klassen and Thompson (1987, 1989, *in press*), and suggests that it may

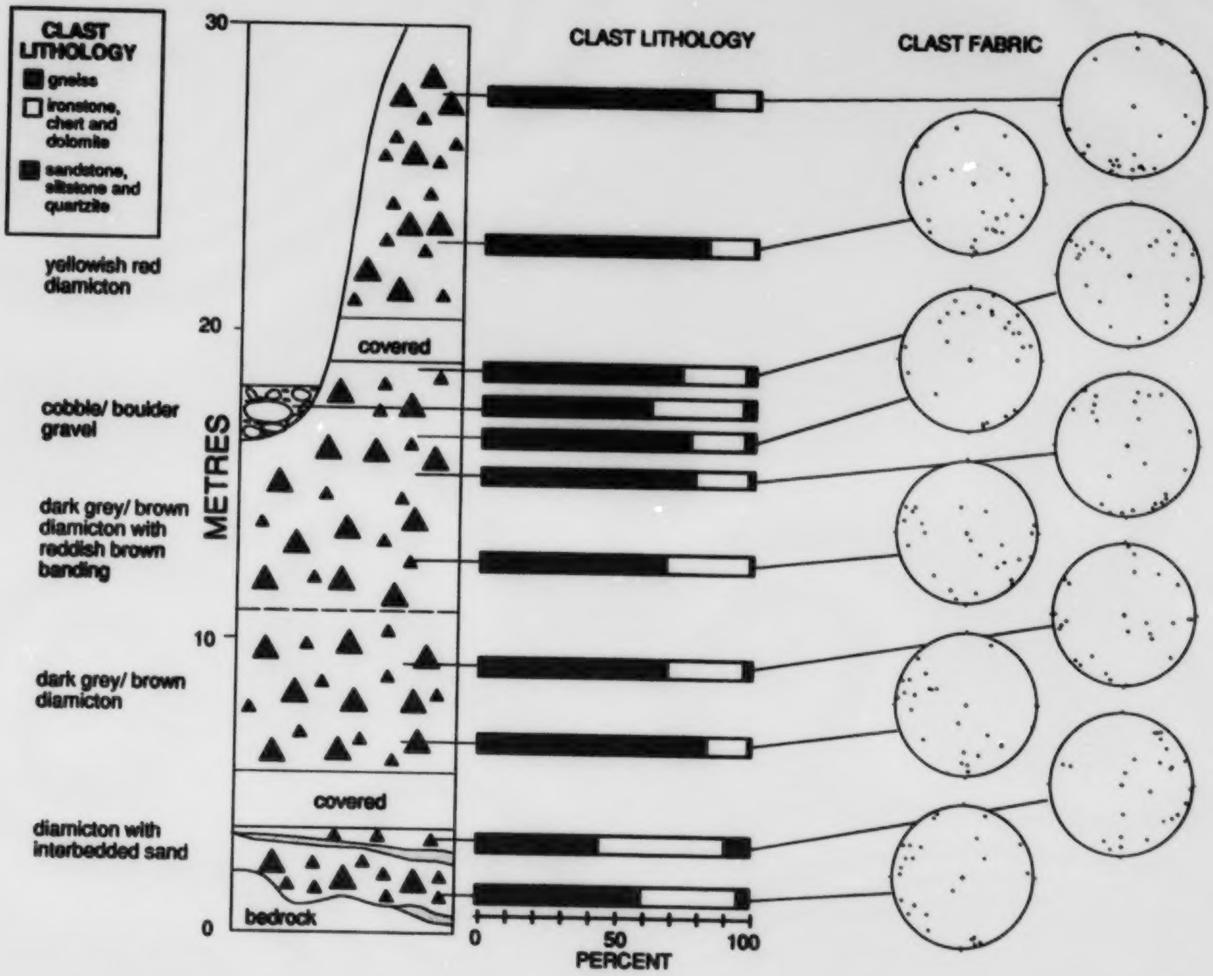
CLAST LITHOLOGY
gneiss
ironstone, chert and dolomite
sandstone, siltstone and quartzite
yellowish red diamicton
cobble/ boulder gravel
dark grey/ brown diamicton with reddish brown banding
dark grey/ brown diamicton
diamicton with interbedded sand
covered
covered
bedrock
METRES
30
20
10
0
CLAST LITHOLOGY
CLAST FABRIC
0
50
100
PERCENT

Figure 8. *Composite section diagram, Burnt Creek 2, showing lithology, clast fabrics and clast lithology.*

find application on a local scale. Considerable local topographic variation from regional patterns is apparent in areas of high relief. Detailed striation mapping at the property level is advisable to provide information regarding local flow conditions. Where well-oriented clast fabrics are found, they suggest northwest–southeast flow, supporting the contention of Kirby (1961) that a later northeast flow had little effect on deposited sediment.

The landscape in the study area is dominantly one of glacial and meltwater erosion, rather than deposition, which results in problems when applying conventional drift-exploration methods. Over much of the area, the effects of meltwater erosion and deposition dominate, with glacial processes having little effect on the surface sediment cover. There is evidence that the meltwater flows affecting the area were able to transport clasts up to boulder size. Therefore, both glacial and glaciofluvial models of dispersal have to be considered. Glacial dispersal models in Labrador have been described by Klassen and Thompson (1989, *in press*) and Batterson (1989a, b). Glaciofluvial models are less well understood, but in general, transport distances should be greater than in glacial systems, and matrix geochemistry harder to interpret (Perttunen, 1989). Flow directions, and likely source directions, can be interpreted from mapping of geomorphology and measurement of orientations of sedimentary structures within the sediments.

Thick sections exposed in mine pits are interpreted as showing the effects of the last glacial event to affect the area, rather than a long glacial history. Exposure in the mine pits was poor, however, and the preservation of soft bedrock suggests that preglacial or interglacial sediments may also be present. The stratigraphy based on diamicton coloration outlined by Klassen and Thompson (1987) was not identified in the present study, and it is suggested here that the surface diamicton shows no consistent colour, but reflects the erodibility and colour of local bedrock.

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GRANULAR AGGREGATE-RESOURCE MAPPING OF THE GRAND FALLS (NTS 2D/13) AND MOUNT PEYTON (NTS 2D/14) MAP AREAS

M.J. Ricketts and T. Christopher
Terrain Sciences Section

ABSTRACT

This report presents preliminary results of 1:50 000-scale granular aggregate-resource mapping in the Grand Falls (NTS 2D/13) and Mount Peyton (NTS 2D/14) map areas of central Newfoundland. The aim of the project is to complete a regional aggregate study to locate suitable deposits for economic extraction by the construction industry. Significant deposits have been located along the banks of the Exploits River, and at Sandy Brook, Stony Brook, West Stony Brook, the Northwest Gander River and near Red Cliff and Diversion Lake.

INTRODUCTION

The 1991 field season was the final year of a 3-year, 1:50 000-scale granular aggregate-resource mapping program in central and northeastern Newfoundland (Figure 1). The objectives of the program were to locate, map and sample sand, gravel and till materials and to evaluate their suitability for use as aggregate. Field sampling concentrated on glaciofluvial outwash deposits, which is one of the major sources of sand and gravel in Newfoundland. These deposits are excellent sources of low silt−clay aggregate, which is required for use in concrete and asphalt and as Class A and B road gravel. Till is generally regarded as poor-quality aggregate due to its high silt−clay content, and therefore, was given less attention during field investigations.

MINERAL AGGREGATE−A REVIEW

Mineral aggregates, as used in the context of this report, are defined as any hard, inert material such as gravel, sand, crushed stone or other mineral material that is used in the construction industry (Carter, 1981). Aggregates are used extensively in all types of construction activities related to domestic, industrial or other developments. Provincially, road construction is a major use of aggregate materials. Municipally, water and sewer systems, driveway construction and building foundations all require aggregate. Backfill is another major use, as is topsoil for landscaping. Aggregate materials can be, (a) used in an unprocessed form as pit and run, (b) processed and used as Class A gravel (i.e., generally aggregate consisting of 100 percent particle sizes less than 19 mm in diameter and 3 to 6 percent silt−clay content, plus a specified range of percentages for size fractions between these upper and lower limits; Department of Transportation, 1987) and Class B gravel (i.e., generally aggregate consisting of 100 percent particle sizes less than 101.6 mm in diameter and containing 3 to 6 percent silt−clay plus a specified range of percentages for size fraction between these upper and lower limits; Department of Transportation, 1987), and (c) mixed with a cementing agent to form concrete, asphalt and mortar.

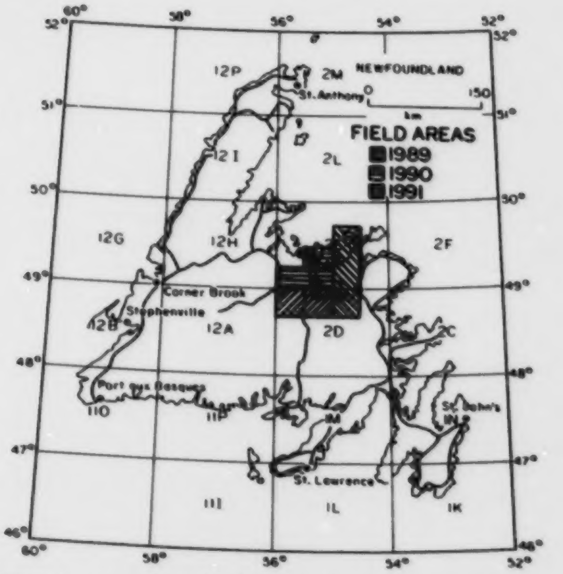

Figure 1. *Location of field area, showing the survey areas for 1989, 1990 and 1991.*

Not all materials are suitable as aggregate. Vanderveer (1983) defined the quality of mineral aggregate by its composition. Aggregates containing too much or too little silt and clay when used in road construction can cause instability, such as flowage in the case of too much fine material, or the loss of compaction properties in the case of too little fine material. Too much fine material in concrete (> 2 percent) can interfere with the bonding process between the aggregate and the cementing agent. The presence of deleterious substances such as silt−clay coating or iron-oxide staining on the surface of the aggregate, or the presence of friable or blade-shape fragments often cause bonding problems with the cementing agent, or the breakdown of aggregate with time.

Aggregates are a high-volume, low-cost material. The cost of transport represents, on average, 30 percent of the delivered price (Vanderveer, 1982). Thus, the location of resources relatively close to the users is important in Newfoundland. The many users throughout the Island makes a regional mapping project of aggregate resources vital.

LOCATION AND ACCESS

Granular aggregate-resource mapping was conducted in the Grand Falls (NTS 2D/13) and Mount Peyton (NTS 2D/14) map areas of central Newfoundland.

Most of the field area is accessible by major road routes, numerous logging roads (in use and abandoned), which are usable by all-terrain vehicle, and water ways, which are navigable by canoe. Foot traversing was required in the south-central parts of both map areas where other means of ground movement was difficult to impossible.

PREVIOUS WORK

Surficial geological mapping at 1:50 000-scale (based mainly on 1978 reconnaissance aerial photographic interpretation) has been completed for NTS 2D/13 and the north and west part of NTS 2D/14 (Kirby *et al.*, 1989).

Reconnaissance aggregate-resource programs conducted by the Department of Mines and Energy (Environmental Geology Section, 1983; Kirby *et al.*, 1983) covered a 6-km-wide corridor along all major roads and a wider radius around towns in the area. Kirby (1991) conducted detailed sampling in granular aggregate deposits along the Exploits River and near Red Cliff. In addition, geotechnical bedrock maps have been compiled at a scale of 1:250 000 (Bragg, 1986). Bedrock aggregate mapping at a scale of 1:50 000 was conducted in central Newfoundland (Bragg, 1990) to locate suitable bedrock deposits as an alternative to granular aggregates.

MAPPING AND ANALYTICAL METHODS

Interpretation of 1:50 000-scale, black and white aerial photographs was used to locate potential deposits of sand and gravel. Following this, interpretation of 1:12 500-scale coloured aerial photographs was conducted in selected areas to delineate deposit boundaries. In these selected deposits where exposures were present, field sampling was carried out at 0.5 to 1.0 km intervals in outwash deposits and at 1.5 to 2.0 km intervals in till. Additional samples were taken in each deposit wherever changes of sediment type were observed, where quality differences were apparent, or wherever sediments of variable quality or texture could potentially be quarried separately.

Where possible, samples were taken from natural exposures such as stream cuts, shorelines, and gullies or from man-made exposures such as road cuts, and pit and quarry excavations. Where these types of exposures were not available, samples were collected from hand-dug pits, 1 m in depth.

Sampling provided material for petrographic and grain-size analyses, and Los Angeles Abrasion (ASTM Standards C 88-83, 1990) and Soundness Tests (ASTM Standard C 131-89, 1990). Field sieving and petrographic analysis were performed on all samples containing >8mm-size material. Approximately 15 kg of material was collected for field sieving. A split of the finer than 8-mm diameter material was retained for laboratory sieve analyses following procedures outlined in Ricketts (1987). Los Angeles Abrasion and Soundness Test results are currently unavailable.

PETROGRAPHIC ANALYSES

Aggregate samples were examined to determine the petrography of the pebble fraction (16 to 32 mm) as a preliminary means of determining aggregate quality for construction purposes. A petrographic number was calculated using procedures similar to those outlined in CSA standard A23.2.30 (Canadian Standards Association, 1973). The petrographic number, which can range from 100 to 1000, is derived by taking the sum of the percentage of each rock type present in the pebble fraction (50 pebbles) multiplied by a petrographic factor (based on soundness and durability) assigned to that rock type (Bragg, 1990). The petrographic factor is determined by the rock type and their grain size in a given pebble sample, and also by the presence of silt–clay coatings, weathering, staining, degree of sphericity, rounding and fractures. The lower the petrographic number, the better the quality of aggregate material. For example, a clean, hard, fresh granite would normally have a petrographic number of 100, whereas a friable, soft shale would be 1000. Most deposits contain a combination of different rock types with different petrographic factors. The proportion of each of these components determines the petrographic number (see Table 1 for example). For most purposes, in Newfoundland, aggregate material used in concrete requires a petrographic number of 135 or less, and in road asphalt and for Class A and B gravels, a petrographic number of 150 or less is acceptable (Department of Transportation, 1987).

Table 1. Example for calculating a petrographic number for a granular aggregate sample having more than one rock type

Rock type	# of pebbles	%		Petrographic factor		Petrographic number
shale	12	22	x	10	=	220
granite	27	50	x	1	=	50
sandstone	10	19	x	3	=	57
schist	5	9	x	6	=	54
Petrographic number for sample analyzed					=	381

RESULTS

A total of 232 sites were examined and 201 samples collected in the NTS 2D/13 map area. They consisted of 92 gravel, 67 till, 38 sand and 2 silt and 2 clay samples. In the

NTS 2D/14 map area, 98 sites were examined and 67 samples collected. Samples consisted of 42 tills, 18 gravels, 6 sands and 1 silt. A computer graph program (available at the Department of Mines and Energy) is used to plot results as cumulative curves (Kirby *et al.*, 1983) and to calculate percentages of gravel, sand, and silt–clay or individual sieve percentages for each sample. Till samples will not be discussed further because of high silt–clay content (>6 percent) and their low economic importance in most industrial projects.

Data, including aggregate-resource maps and grain-size analyses, will become available through open-file release.

POTENTIAL DEPOSITS

The following are brief descriptions of the main granular aggregate deposits in the study area (Figure 2; Table 2). The descriptions include deposit location, dimensions, average percentage of gravel, sand, and silt–clay, and the range of petrographic numbers for each deposit. Results of particle size and petrographic analyses are based on a compilation of sample data collected during the 1991 field season and previous sampling programs by the Department of Mines and Energy.

MAP AREA NTS 2D/13

1. NORTHWEST EXPLOITS RIVER

The northwest Exploits River deposit is located in the west of map area NTS 2D/13 on the north bank of the Exploits River (Figure 2). It is approximately 3 km long and has an average width of 0.3 km. Kirby (1991) subdivided this deposit into two zones based on terrain type; zone A is a glaciofluvial terrace and is estimated to be approximately 6 to 8 m thick; zone B is an eroded terrace consisting of several gravel ridges and abandoned dry river channels cut by higher levels of the Exploits River during glacial meltwater run off. These ridges are estimated to be 10 to 15 m thick.

Grain-size analyses of sampled material indicate both zones are moderately sorted and contain high gravel content, moderate amounts of sand and low percentages of fines (Table 2). Two samples in zone A contain silt–clay in the 9 and 10 percent range (Figure 3) but this is a minor problem considering the overall size of the deposit. Most samples collected from this deposit were from backhoe-dug pits and supplemented by hand-dug pits and river-bank exposures. Petrographic numbers of the 16- to 32-mm pebble fraction generally have high values. Kirby (1991) attributes the poor petrographic quality to moderately weathered conglomerate pebbles and to a lesser extent, weathered acidic volcanic pebbles.

Based on grain-size distributions and volume this deposit has a high potential for quarry development. However, petrographic characteristics indicate this material is unsuitable as a high-quality aggregate source.

2. ASPEN BROOK–LEECH BROOK

The Aspen Brook–Leech Brook deposit is located along the north bank of the Exploits River between Aspen Brook and Leech Brook in the western part of map area NTS 2D/13 (Figure 2). The deposit is 8 km long, has a width between 0.2 km and 0.8 km, and a thickness of 2 to 9 m increasing eastward toward Leech Brook. The Leech Brook area has been the site of a large quarry operation in the past (Plate 1). It is a coarse-grained, moderately sorted deposit (Table 2). Three samples collected with silt–clay contents of 10.7, 12.7 and 12.8 percent (Figure 4), are not typical of material in the area and were not included in average analyses. Petrographic numbers range from good to poor in this deposit based on the analyses of 14 samples. Samples with poor petrographic quality are the result of weathered siltstone and sandstone. There were also a number of highly weathered undefined pebbles, which contributed significantly to poor petrographic quality in some samples.

Due to the past removal and a thin gravel unit in parts of this deposit, remaining quarry potential is considered low to moderate.

3. SOUTHWEST EXPLOITS RIVER

The southwest Exploits River deposit is a partly dissected glaciofluvial terrace located at the west end of map area NTS 2D/13 on the south bank of the Exploits River (Figure 2). It is one of the largest deposits in the study area, beginning 3 km west of Tom Joe Brook and extending for 14 km eastward along the Exploits River to Sandy Brook. It has an average width of approximately 0.7 km, and thicknesses of between 3 and 10 m. Grain-size data of 94 samples indicate this deposit is variably sorted and contains equal amounts of sand and gravel, and generally has less than 1 percent silt–clay (Table 2). Eleven other samples had silt–clay contents between 10 and 52 percent (Figure 5). Some of these samples having high silt–clay concentrations were taken from thin units within gravel beds or from sites away from the larger gravel sources within the deposit. Due to the extensive area covered by this deposit, silt–clay concentrations should not be a major concern if quarrying occurs. Petrographic numbers are generally high throughout the deposit. Poor petrographic quality is due to 25 percent, slightly weathered pebbles consisting of siltstone, sandstone and some volcanic rocks.

Based on grain-size analyses and volume, the southwest Exploits River deposit is considered to have high potential for quarry development. However, high petrographic numbers reduce its potential for a high-quality aggregate source.

4. RED CLIFF

The Red Cliff deposit is situated on the north side of the Exploits River, 1.5 km west of Rushy Pond (Figure 2). It is composed of a glaciofluvial terrace and esker complex. The Red Cliff deposit can be divided into 3 zones, based on the grain-size distribution (Table 2; Figure 6) of 22 samples collected in quarry and road-cut exposures.

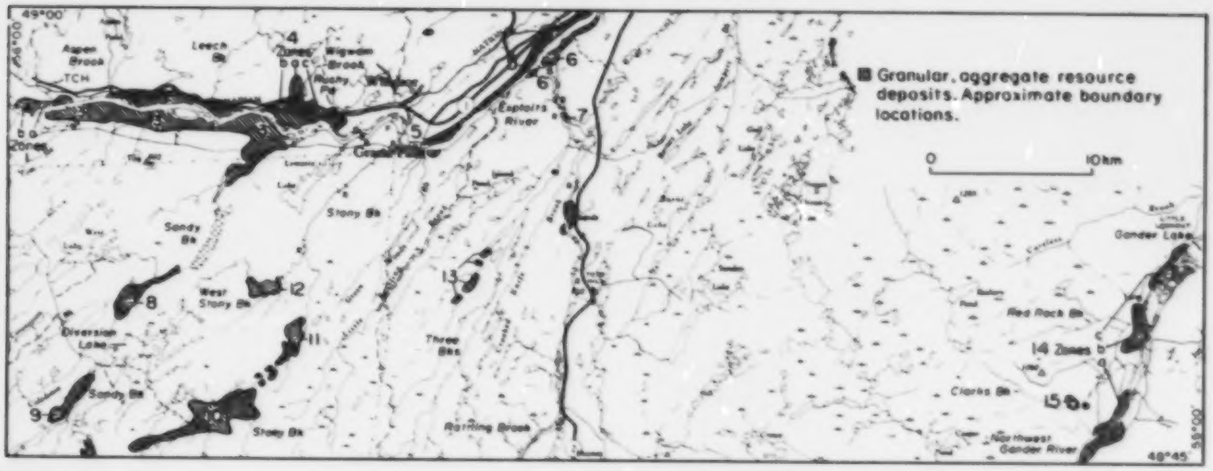
49°00'
56°00'
Aspen Brook
Leech Bk
TCH
4
Zones b c
Wigwam Brook
Rushy Pd
6
7
5
Exploits River
Grand Falls
b a
Zones
Stony Bk
Sandy Bk
8
West Stony Bk
12
13
11
10
Diversion Lake
Sandy Bk
9
Stony Bk
Three Bks
Rattling Brook
Granular, aggregate resource deposits. Approximate boundary locations.
0
10km
Gander Lake
Red Rock Bk
14 Zones b
a
c
Clarks Bk
15
Northwest Gander River
48°45'
55°00'

Figure 2. *Potential aggregate-resource deposits in the Grand Falls (NTS 2D/13) and Mount Peyton (NTS 2D/14) map area.*

Table 2. Summary and comparisons of aggregate deposits in the NTS 2D/13, 14 map areas of central Newfoundland

		Grain-size Analyses				Petrographic Analyses		
Deposit	Estimated m3 x 6	# of samples analyzed	% Grv. (+5mm)	% Sd. (+78um to 5mm	% Slt-Cly (-78um)	# of samples analyzed	Range	Average
1. Northwest Exploits River								
Zone A	>10	15	69.0	30.5	0.5	15	154-253	212
Zone B	>10	27	69.7	28.7	1.5	27	128-262	193
2. Aspec Brook–Leech Brook	1.0-5.0	14	74.5	25.3	0.3	14	100-266	175
3. Southwest Exploits River	>10	94	49.9	49.3	0.7	50	116-653	282
4. Red Cliff								
Zone A	1.0-5.0	14	41.1	56.1	2.9	6	100-235	149
Zone B	1.0-5.0	3	52.6	47.2	0.3	1	148-148	148
Zone C	1.0-5.0	5	0.7	90.8	8.5	0	–	–
5. Northwest Bank of Exploits River	1.0-5.0	14	46.3	52.8	0.9	9	212-315	267
6. Rattling Brook–Exploits River	1.0-5.0	9	58.7	41.1	0.2	8	230-919	448
7. Rattling Brook	<1.0	6	31.0	66.0	3.0	2	314-380	348
8. Sandy Brook	>10	6	73.1	26.7	0.2	6	201-414	294
9. Diversion Lake	>10	7	46.6	50.4	3.1	4	246-334	289
10. Stony Brook South	>10	6	51.0	47.6	1.4	4	176-408	242
11. Stony Brook North	1.0-5.0	5	61.1	37.2	1.8	4	209-286	260
12. West Stony Brook	1.0-5.0	4	23.6	75.7	0.7	1	260	260
13. Three Brooks	<1.0	13	51.5	48.3	0.2	11	224-506	315
14. Northwest Gander River								
Zone A	5.0-10	12	40.2	59.3	0.6	8	130-352	237
Zone B	1.0-5.0	3	78.5	21.3	0.2	2	128-158	143
Zone C	1.0-5.0	4	59.7	39.8	0.5	3	130-350	215
15. Charles Brook	<1.0	2	72.7	27.1	0.2	2	112-150	131

Note: Estimated quantities in Table 2 are based on air-photo analysis and field investigation of road cuts, shallow hand-dug pits and natural exposures. Detailed sampling methods such as drilling were not conducted. Grain-size analyses are based on sample data averages for each deposit and do not take into account extent and thickness of units at any one location.

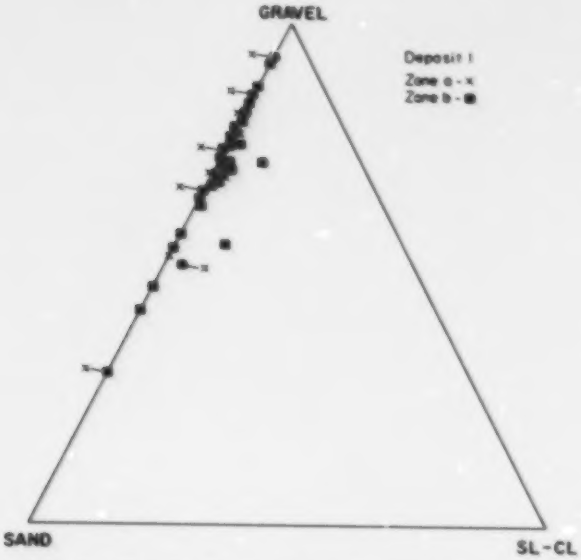

Figure 3. *Ternary diagram showing granular distribution of samples collected in the northwest Exploits River deposit.*

Plate 1. *Inactive quarry in the Aspen Brook–Leech Brook deposit.*

Zone A consists of an esker complex located north of the Trans Canada Highway on the west side of Wigwam Brook, and is the largest of the three zones. Grain-size analyses of samples indicate that this area is composed of a sandy-gravel, which has a low silt–clay content. Stratified units in quarry exposures show poorly sorted gravel and moderately to well-sorted sand. Petrographic analyses of sampled material range from very good to poor. High petrographic numbers are the result of weathered sandstone and siltstone. This zone has been extensively quarried in the past, but has a considerable amount of material remaining for future quarry activity. However, consideration should be given to poor petrographic quality, when considering this material for concrete and road-paving projects.

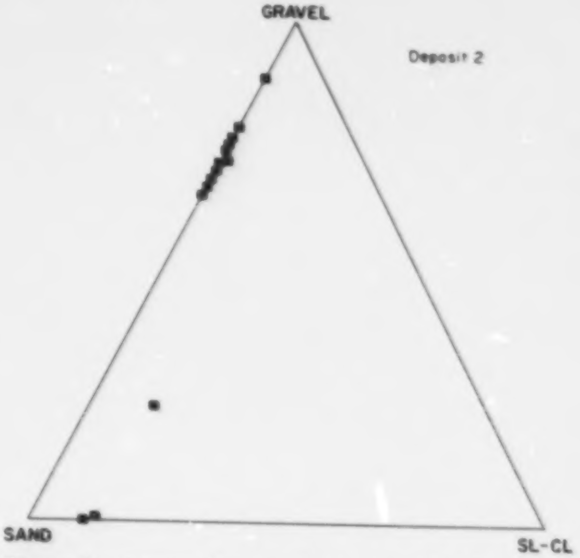

Figure 4. *Ternary diagram showing granular distribution of samples collected in Aspen Brook–Leech Brook deposit.*

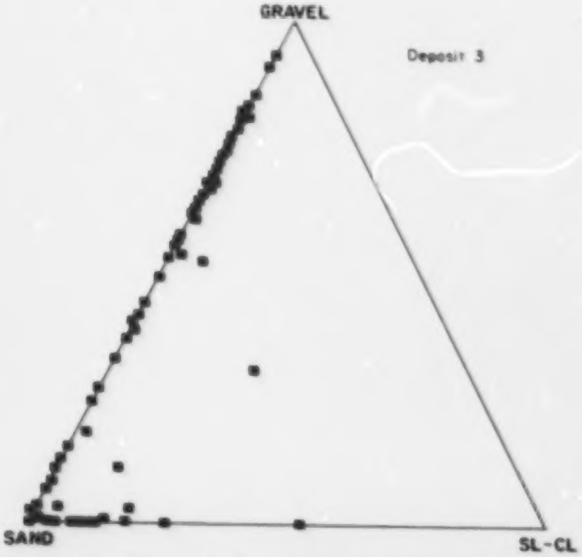

Figure 5. *Ternary diagram showing granular distribution of samples collected in the southwest Exploits River deposit.*

Zone B is located along the north side of the Exploits River, south of zone A. It is a thin outwash deposit (generally < 3 m thick) consisting of poorly sorted sandy gravel generally greater than 0.125 mm in diameter (97.5 percent). A petrographic number of 148 was determined from the petrographic analyses of one sample. Petrographic quality of this zone is reduced by the presence of moderately weathered to weathered siltstone. Past quarry activity in this area has nearly depleted the resource. It has a low potential for future quarry development.

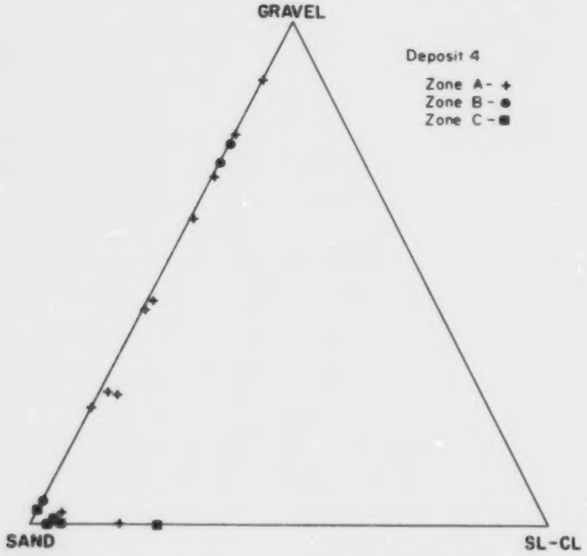

Figure 6. *Ternary diagram showing granular distribution of samples collected in the Red Cliff deposit.*

Zone C is a moderately to well-sorted sand deposit located between zones A and B, extending east between Rushy Pond and the Exploits River. This zone is estimated to be between 3 and 5 m thick but may locally reach thicknesses of 8 m. It contains in excess of 90 percent sand (grain-size diameter is generally less than 0.5 mm) with the silt–clay content ranging from 0.5 to 24.4 percent in 5 samples analyzed. Due to the fine texture of this deposit there were no pebbles to conduct petrographic analyses. Based on volume alone, the potential for a sand quarry in this area is high, however, the high silt–clay content and fine sand texture may reduce its usefulness for most construction purposes.

5. NORTHEAST BANK OF EXPLOITS RIVER

The northeast bank of the Exploits River in the northeast corner of map area NTS 2D/13 consists of a glaciofluvial terrace deposit over 12 km long and 0.1 km to 0.7 km wide (Figure 2). It is generally 6 to 8 m thick, becoming thinner (approximately 2 to 3 m) in the northeast, past the confluence with Rattling Brook. Grain-size analyses of 14 samples indicate that this deposit is a poorly sorted gravelly sand, having low silt–clay content (Table 2; Figure 7). Silt–clay samples (not included in average analyses) were collected at the base of two exposures of this deposit. Petrographic numbers of 9 samples analyzed were high. Poor petrographic quality is principally due to moderate to highly weathered micaceous sandstone and weathered siltstone.

This deposit has moderate to low potential for quarry development. Silt–clay units may pose problems in some areas and the proximity of a large section of the deposit to the Exploits River will restrict exploitation. High petrographic numbers will reduce its potential as a high-quality aggregate for most construction purposes.

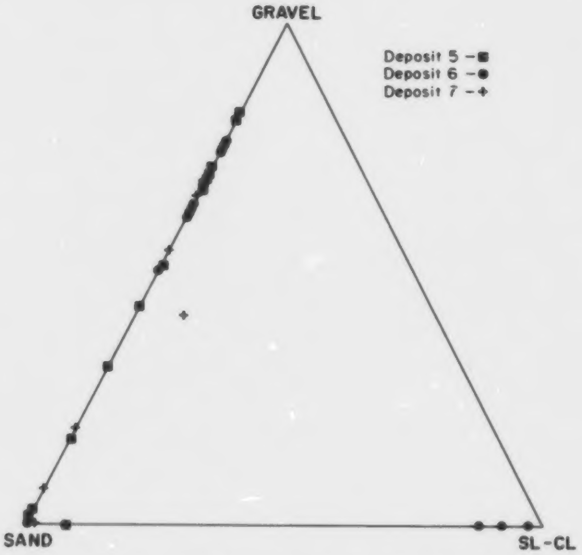

Figure 7. *Ternary diagram of granular distribution of samples collected in deposits 5, 6 and 7 along the northeast bank of the Exploits River, confluence of Rattling Brook and Exploits River and along the east bank of Rattling Brook, respectively.*

6. RATTLING BROOK–EXPLOITS RIVER

The Rattling Brook–Exploits River deposit is a glaciofluvial terrace located in the northeast part of map area NTS 2D/13 at the confluence with Rattling Brook and the Exploits River (Figure 2). This deposit is over 6 km long and 0.1 to 0.6 km wide. Thickness varies from 1 m in the northeast to 8 m in the southwest. This deposit is well stratified and contains units of gravel, sandy gravel and silt–clay (Table 2). Fine-grain units of silt–clay were sampled in 3 exposures below gravel beds. These fine-grain sediments (Figure 7) are located below the gravel units and were not included in the average analyses. Poor petrographic quality in this deposit results principally from moderate to highly weathered sandstone and siltstone.

The Rattling Brook–Exploits River deposit has a moderate to low potential for quarry development. Thin gravel beds, zones of silt–clay, past removal, private land ownership and poor petrographic quality will have a negative influence on quarry development in this area.

7. RATTLING BROOK

The Rattling Brook deposit is a small glaciofluvial terrace situated on the east bank of Rattling Brook, 3.5 km up stream from its confluence with the Exploits River (Figure 2). This deposit has an average length of 1 km and and average width of 0.7 km. River-bank exposures (Plate 2) show an average thickness of 3 m. It consists of stratified, poorly sorted gravels and moderately sorted sand with generally low silt–clay contents (Table 2). Five of the 6 samples have a range from 0.4 to 1.1 percent silt–clay and the sixth sample has 16 percent silt–clay (Figure 7). Petrographic quality in this deposit is poor because of shale content and moderate to highly weathered sandstone.

Plate 2. *A 3-m-high river-bank exposure along Rattling Brook.*

The Rattling Brook deposit has a poor potential for quarry development. Shallow thickness, private land ownership and poor petrographic quality will negatively affect development in this area.

8. SANDY BROOK

The Sandy Brook deposit is located along Sandy Brook in the west part of map area NTS 2D/13 (Figure 2). This deposit is over 5 km long and 0.1 to 1 km wide. It is composed of a glaciofluvial terrace and two eskers. The terrace covers most of the deposit area but has a thickness of only 2 to 4 m. The eskers are 0.7 km and 1 km long, respectively, and are estimated to be 12 to 15 m thick. No natural exposures were found in this deposit and therefore grain-size analyses is based on samples taken from hand-dug pits. Both esker and terrace deposits contain a large amount of gravel, lower amounts of sand and have a low silt–clay content (Table 2; Figure 8). Petrographic quality is poor, resulting from weathered sandstone and shale pebbles.

The Sandy Brook deposit has a moderate to high potential for quarry development based on granular content and potential reserves of material, although poor petrographic quality reduce its potential as a high-quality aggregate for use in most construction projects.

9. DIVERSION LAKE

The Diversion Lake deposit is situated 1 km southwest of Diversion Lake in the southwest part of map area NTS 2D/13 (Figure 2). It is a hummocky glaciofluvial outwash deposit 4 km long and averaging 0.8 km wide. Thickness varies between 4 and 7 m. It is a sand–gravel deposit having low silt–clay content (Table 2). Two of seven samples collected in this deposit had silt–clay contents of 6.8 and 11.5 percent, respectively (Figure 8). Five other samples had less than 1.0 percent silt–clay. Petrographic quality in this deposit is poor, resulting from weathered sandstone, tuff, basalt and gabbro pebbles.

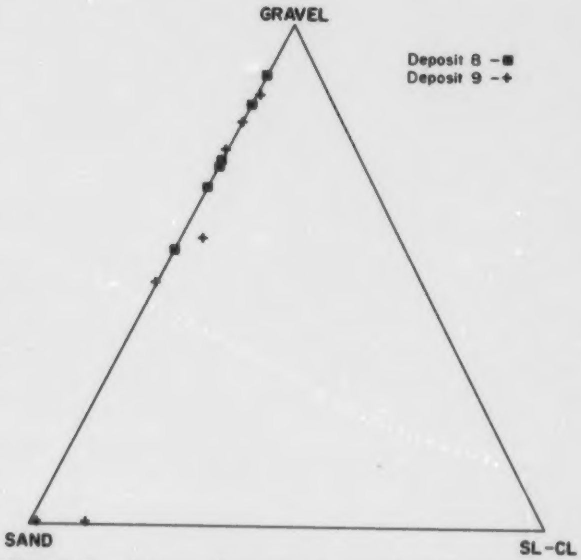

Figure 8. *Ternary diagram of granular distribution of samples collected in the Sandy Brook and Diversion Lake; deposits 8 and 9, respectively.*

Sediment in the Diversion Lake deposit was used in the past for construction of forest access roads. This deposit has a high potential for future quarry development based on volume and grain-size distribution. However, poor petrographic quality reduces its potential for concrete and most road-construction projects.

10. STONY BROOK SOUTH

The Stony Brook south deposit, located along Stony Brook in map area NTS 2D/13 (Figure 2), consists of a large glaciofluvial terrace over 8 km long having an average width of 1 km. River-bank exposures (Plate 3) observed during field investigations indicate that thicknesses of 6 m to 12 m is common. It is a poorly sorted gravel and sand deposit having low silt–clay content (Table 2; Figure 9). Petrographic quality is poor and variable, resulting from weathered sandstone, siltstone and a few weathered, granite pebbles.

This is the largest deposit in the study area and has a high potential for quarry development based on potential reserves and grain size. However, variable petrographic quality indicates a need for more extensive sampling and analyses to determine use for this material as a high-quality aggregate.

Plate 3. *A 6-m-high river-bank exposure in the Stony Brook south deposit.*

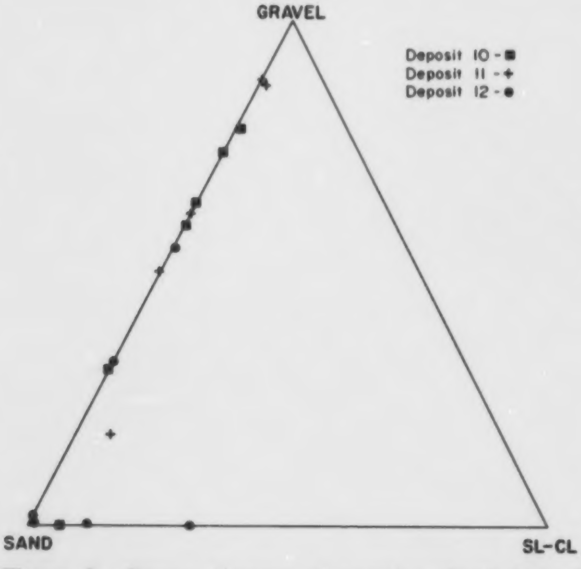

Figure 9. *Ternary diagram of granular distribution of samples collected in deposits along Stony Brook south and north and West Stony Brook; deposits 10, 11 and 12, respectively.*

11. STONY BROOK NORTH

The Stony Brook north deposit is located on the northwest bank of Stony Brook in the south-central part of map area NTS 2D/13 (Figure 2) and consists of an esker ridge complex (Plate 4). Glaciofluvial outwash sediment is found in adjoining areas and in thin layers between the esker ridges. The esker complex is 1.4 km long and 0.6 km wide and is generally less than 5 m thick. Grain-size analyses of samples collected from this (poorly sorted) deposit show it is composed of a sandy gravel with low silt−clay content (Table 2). Four of five samples have < 1 percent silt−clay and the other sample has 7.5 percent silt−clay (Figure 9). Petrographic analyses indicate poor-quality material, resulting from weathered siltstone and sandstone. Some weathered basalt and tuff clasts are also present.

The Stony Brook North deposit has a moderate to low potential for development because of small aggregate volumes and poor petrographic quality.

Plate 4. *Road-cut through esker ridge in the Stony Brook north deposit.*

12. WEST STONY BROOK

The West Stony Brook deposit is situated along the south side of West Stony Brook in the central part of map area NTS 2D/13 (Figure 2). This is a glaciofluvial terrace deposit, approximately 1.7 km long, having an average width of 0.7 km. Grain-size analyses of four samples collected from quarry exposures (Plate 5) and road cuts indicate this deposit is a moderately to poorly sorted sand with low amounts of gravel and silt−clay (Table 2). Two other samples with silt−clay contents of 11.8 and 31.2 percent (Figure 9) were not included in average analyses. These sites appeared to be localized and should not affect quarry operations in other parts of the deposit. Petrographic analysis was conducted on only one sample due to the fine texture in most of the deposit. It produced a petrographic number of 260, the poor petrographic quality being principally the result of weathered volcanic rocks.

The West Stony Brook deposit has a high potential for a sand quarry operation, although further sampling should be done to provide a broader range of petrographic data.

Plate 5. *Abandoned quarry in the West Stony Brook deposit.*

13. THREE BROOKS

The Three Brooks deposit is situated on the east side of map area NTS 2D/13. It consists of four small glaciofluvial terraces, accessible along a 4 km section of forest access road. The largest terrace is approximately 1 km long, 10 m wide and 4 to 6 m thick. The other three terraces are approximately 400 m long, 300 m wide and 2 to 4 m thick. Samples collected from backhoe-dug pits, stream cuts (Plate 6) and road cuts indicate that all four areas consist of poorly sorted sandy gravel and low silt–clay concentrations (Table 2; Figure 10). Petrographic quality is poor resulting from weathered sandstone, siltstone and shale. Weathered volcanic rocks, granite, gabbro and conglomerate also contribute to the poor quality.

Plate 6. *A 3-m-high river-bank exposure in the Three Brooks deposit.*

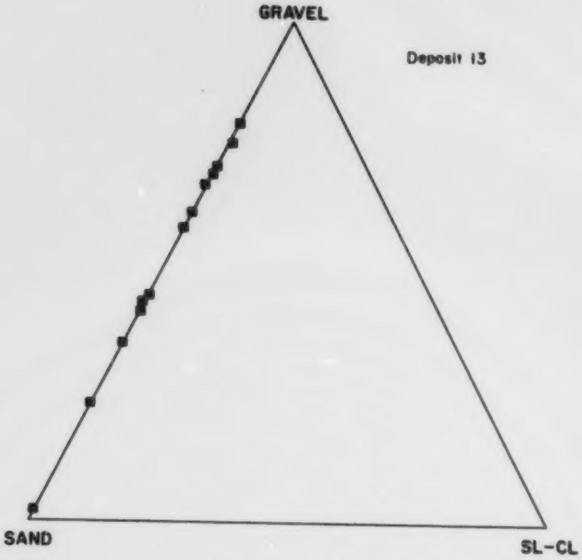

Figure 10. *Ternary diagram of granular distribution of samples collected in the Three Brooks deposit.*

Potential quarry development in this area is considered moderate in the largest of 4 terraces, and low in the remaining 3 areas where gravel beds are thin. However, poor petrographic quality reduces its potential as a high-quality aggregate.

MAP AREA NTS 2D/14

14. NORTHWEST GANDER RIVER

The banks of the Northwest Gander River located in the southeast corner of map area NTS 2D/14 (Figure 2) consist of glaciofluvial terrace sediment that can be divided into three different zones based on grain size (Table 2; Figure 11).

Zone A is the most southerly and the largest of the three areas. It is approximately 4 km long and 0.3 km wide, and varies in thickness from 2 to 8 m. This area consists of poorly sorted sandy gravel and moderately sorted sand, both with low silt–clay content (Figure 11). Petrographic quality is generally poor, resulting from moderately weathered sandstone and siltstone and highly weathered undefined pebbles.

Zone A along the Northwest Gander River has been used for construction of forest access roads, but has sufficient reserves to be considered as a high-potential site for future quarry development. However, variable petrographic quality indicates a need for more sampling and analyses in this area.

Zone B is located on the west side of the Northwest Gander River, south of Red Rock Brook. It is approximately 1 km long and 1 km wide. Road-cut exposures (Plate 7) indicate it is approximately 3 m thick but may locally reach a thickness of 7 m. Zone B is a poorly to moderately sorted gravel with low sand and silt–clay contents (Figure 11). Petrographic numbers of 128 and 158 were calculated from

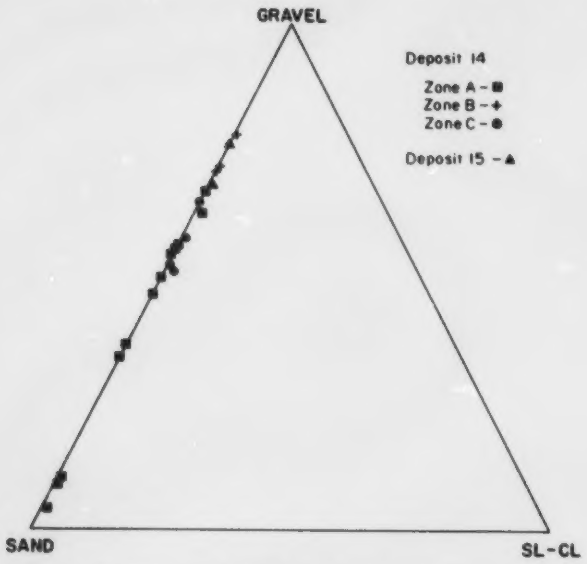

Figure 11. *Ternary diagram of granular distribution of samples collected in the Northwest Gander River and Clarks Brook; deposits 14 and 15 respectively.*

Plate 7. *A 3 m road-cut in zone B along the Northwest Gander River.*

the analyses of two samples. Pebbles consist mainly of fresh granite.

Although the site area is widespread, it is generally thin and quarry operations in this zone will require a large working area. A large amount of topsoil, trees and root debris will have to be removed thereby increasing cost of quarry operation. As a result, this deposit is considered to have a moderate potential for quarry development.

Zone C is located at the outlet of the Northwest Gander River into Gander Lake. It is 3.5 km long and varies from 50 to 350 m wide. Thicknesses range from 3 to 8 m. This is a poorly sorted sandy gravel deposit with low silt–clay content (Figure 11). Petrographic quality ranges from good to poor based on analyses of three samples. The quality of the deposit is reduced by the presence of weathered sandstone and some highly weathered undefined pebbles.

This last zone in deposit 14 has a moderate to low potential for quarry development. Areas of thin sediment and private land ownership will have a negative influence on quarry development in this area, and the variable petrographic quality indicates a need for more sampling and analyses.

15. CLARKS BROOK

The Clarks Brook deposit is an esker complex situated in the southeast corner of map area NTS 2D/14 (Figure 2). It is approximately 500 m long, 200 m wide and 3 to 8 m thick. Grain-size analyses of 2 samples collected from a hand-dug pit and a stream cut indicate this deposit contains poorly sorted gravel with low silt–clay content (Table 2; Figure 11). Petrographic analyses of two samples are 112 and 150, indicating a good-quality aggregate.

Based on potential reserves and grain-size data, the Clarks Brook deposit may have a high potential for quarry development.

OTHER DEPOSITS

Resource areas not numbered in Figure 2 also contain sand and gravel. However, these deposits are small, with depleted or low reserves, restricting continued quarry operations or new quarry development.

SUMMARY

Granular deposits were sampled within the study area to determine their potential for use in the aggregate industry. Several deposits are thin (around 2 to 3 m) and would require extensive topsoil, vegetation and root-debris removal to mine small amounts of material, thereby increasing excavating cost. Deposits of this type are located along Stony Brook, Rattling Brook, near Three Brooks, zones A and B along the Northwest Gander River, zone B at Red Cliff, and parts of deposits along the Exploits River.

Many of the deposits sampled have sufficient quantities of material to support large quarrying operations (exceeding a 10-year production period). These include the northwest and southwest Exploits River deposits, zone A at Red Cliff, the Sandy Brook, Diversion Lake, West Stony Brook deposits, and zone A along the Northwest Gander River.

Low silt–clay content is necessary for aggregates used in concrete projects and asphalt. Most deposits sampled have a silt–clay content of less than 1 percent. Other deposits may have zones of sand–silt or silt–clay, which can be avoided during mining. This reduces the need for washing the material, thereby reducing overall cost. In some areas, screening or crushing may be necessary to remove the coarse aggregate fraction.

Results of petrographic analyses indicate generally poor petrographic quality throughout the study area. Samples from four deposits show petrographic numbers low enough to indicate suitable aggregate material for use in concrete or asphalt type projects. These deposits are located along the northwest Exploits River, the Aspen Brook–Leech Brook deposit, zone A at Red Cliff, along the Northwest Gander River and at Clarks Brook. Within some of these deposits, samples with higher petrographic numbers were also collected. To determine the suitability of aggregates for most concrete projects more extensive testing is needed other than petrographic analyses, as outlined by the CSA Standards for concrete materials. If granular materials are found to be unsuitable, alternative aggregate sources from outside the area or suitable bedrock sources closer to the construction site should be used.

ACKNOWLEDGMENTS

Paul Miles is thanked for locating suitable field accommodations in the Grand Falls area. Earlier mapping by Fred Kirby and Randy Ricketts of the Department of Mines and Energy provided valuable information for planning this project. Emmanual Goosney is thanked for his dedication in operating a skidder-mounted backhoe to dig test pits during the long work days and sometimes intense rainfall in late July. Improvements from critical reviews of this manuscript by David Liverman and Martin Batterson are greatly appreciated.

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ICE FLOW IN WEST-CENTRAL NEWFOUNDLAND

L. St. Croix and D. Taylor
Terrain Sciences Section

ABSTRACT

Reconnaissance-level striation mapping in west-central Newfoundland covering map areas NTS 2D/13, 2E/4, 2E/5, 2E/12 and 12H/1 suggests the existence of four separate ice-flow events. Of these, the earliest ice-flow direction was northeastward, the second was northward and the third ice flow identified had a north-northeastward trend. The fourth and most recent ice flow had an eastward trend. This recognition of the area's recent glacial history, in terms of ice-striae characteristics, will assist in the exploration efforts for gold and base metals.

Exploration activity is generally hampered by the extensive, sediment overburden and lack of bedrock exposures. Studies such as this, which improve our understanding of an area's recent glacial history, enhances the efficiency and effectiveness of drift-prospecting programs.

INTRODUCTION AND OBJECTIVE

West-central Newfoundland is an area of considerable mineral-exploration interest, especially for gold and base metals. A thorough understanding of the area's recent ice-flow and glacial history could help to enhance the efficiency and effectiveness of exploration in drift-covered areas.

The project objective is to compile reconnaissance-level maps of glacial striae, at a 1:50 000 scale, of map areas NTS 2D/13, 2E/4, 2E/5, 2E/12 and 12H/1 (Figure 1).

During the 1991 field season, ice-flow data were collected from west-central Newfoundland; this data gathering and compilation is part of an on-going regional striation study. The purpose of this study is to improve our understanding of the deglacial history of Newfoundland. A tentative model of deglaciation for the Notre Dame Bay area was presented by St. Croix and Taylor (1991). The data collected during this past field season support the sequence of deglaciation proposed by the model and also extend the western margin of the model (see St. Croix and Taylor, 1991) to the eastern edge of the Topsails.

A complete listing of the striation sites recorded during the past summer can be obtained from the Striation Database (Taylor *et al.*, 1991). Efforts continue on the compilation and development of this database for the Island of Newfoundland. To date, approximately 5000 striations sites have been entered into the database. It is planned to extend this coverage over the coming year and to concentrate on specific problems and areas of less than adequate regional coverage.

RESULTS

Approximately 500 single- and multiple-striation sites were observed and recorded during the field season. Ice-flow direction was determined from the morphology of the bedrock surface, using such features as stoss-and-lee, chattermarks and miniature crag and tails. Sites that indicate multiple ice-flow directions commonly indicate separate flow events, but in some cases may represent a different ice-flow direction within one flow event. Relative temporal relationships between the flows are indicated by crosscutting characteristics and leeside preservation.

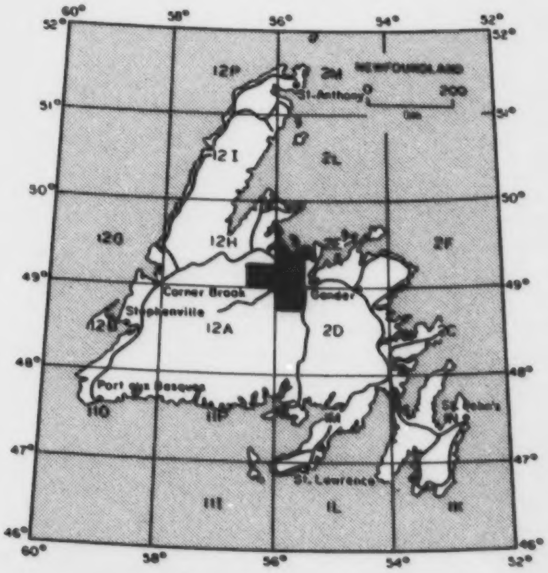
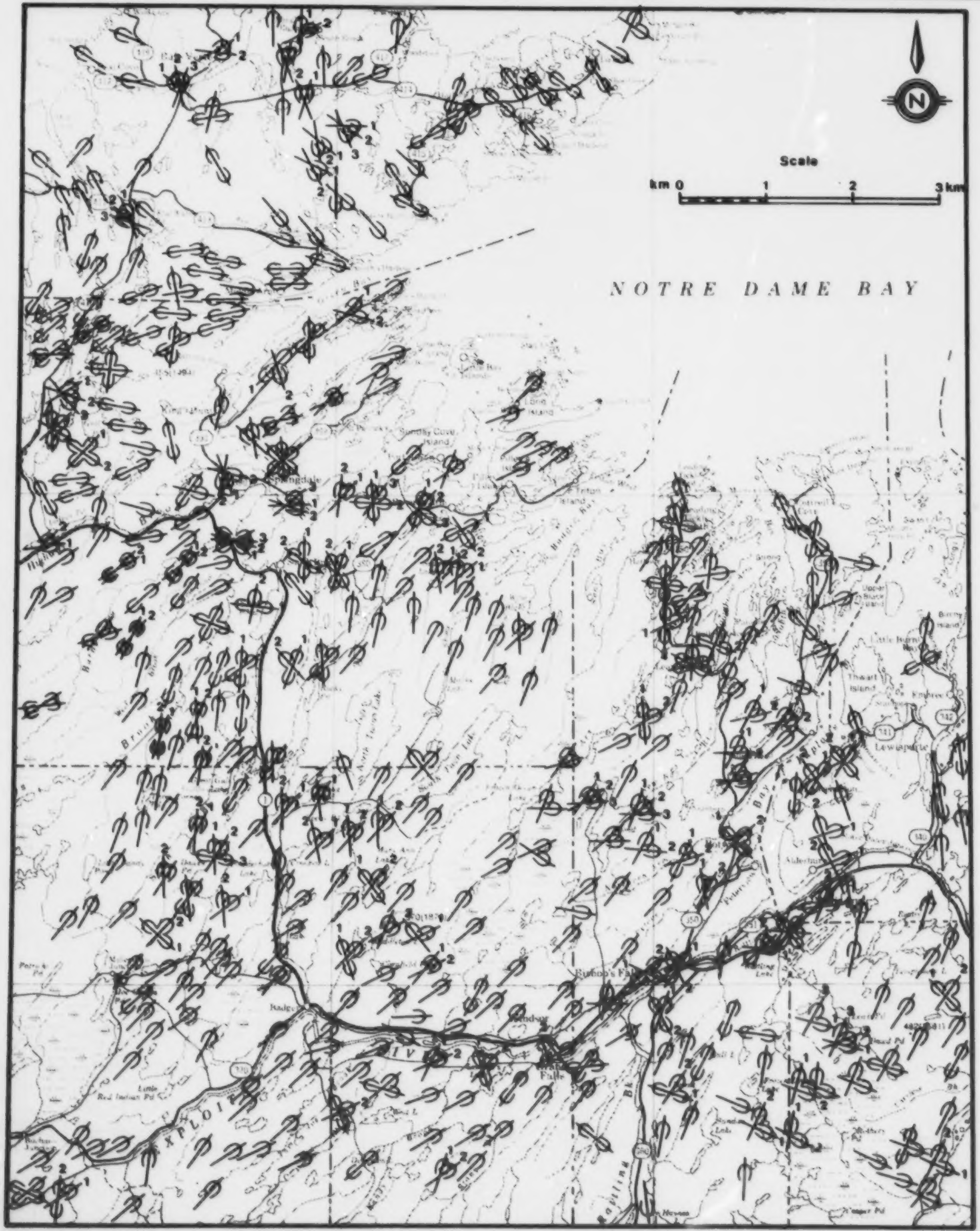


Figure 1. *Map showing the study area.*

The striation data collected suggests the existence of four separate ice-flow events over all or parts of west-central Newfoundland (Figure 2). Of these, the earliest ice-flow


N
Scale
km 0
1
2
3 km
NOTRE DAME BAY
Springdale
Lewisporte
Bishop's Falls
Thwart Island
Little Red Indian Pd

Glacial striae (direction known, unknown). Number indicates relative age, 1 being the oldest.

Figure 2. *Ice-flow in west-central Newfoundland. Figure includes data from: Liverman and St. Croix (1989a), Liverman and Scott (1990), Liverman* et al. *(1991), Sparkes (1984), Taylor and St. Croix (1989, 1990) and Vanderveer and Sparkes (1980).*

direction trends northeastward and was found mainly in the southern half of the field area; the second ice-flow direction is northward and is identified at several sites throughout the region; the third ice-flow has a north-northeastward trend and is also widespread over the entire region. The fourth and most recent ice-flow direction has a eastward trend and is found in the central portion of the field area in the vicinity of New Bay Pond. The absolute age of these ice-flow events is not known.

A northeastward ice flow is the earliest of the four glacial events recorded. Ice-flow indicators that pertain to this ice-flow event have consistently been preserved on the leeside surfaces of outcrops that show one or more subsequent ice-flow events. This flow is believed to have originated from an ice centre in the Red Indian Lake area, and can be found across most of north-central Newfoundland (St. Croix and Taylor, 1991).

The second glacial event, a northward ice flow, is found at various localities throughout the study area. The broad coverage and consistency of this ice-flow event indicate a possible ice divide, directly south of the study area (St. Croix and Taylor, 1991), with northward flow into Halls Bay, Badger Bay and Seal Bay.

The third ice flow, north-northeastward, is again widespread over the whole area. This ice flow consistently crosscuts the earlier (second) northward flow where observed on the same outcrops. This ice flow is generally interpreted to be the last regional ice-flow event in the area, and is commonly the only ice flow represented on many of the outcrops observed.

The most recent ice flow in the region was a relatively local, late-glacial event. This eastward ice flow was identified in the New Bay Pond area and may be related to a remnant ice centre in the Twin Ponds area north of Grand Falls as described by Grant (1974). This final ice-flow event may be related to a cooling episode following initial deglaciation identified in north-central Newfoundland by Eyles (1977), Macpherson and Anderson (1985) and Liverman and St. Croix (1989b).

IMPLICATIONS FOR DRIFT PROSPECTING

Explorationists working in drift-covered areas such as west-central Newfoundland must consider multiple ice-flow events when collecting and interpreting data. A geochemical or sediment dispersal pattern in till is commonly related to the most recent ice-flow event. However, in areas where multiple ice flows have been identified, the earlier ice flows should be taken into consideration because they may also have dispersed sediment along their respective flow paths, which may have been reincorporated by later ice-flow events.

In areas of multiple ice-flow events, preservation of multiple till units is also possible. Each till unit show sediment-dispersal patterns related to the ice flow that deposited it. Thus, an ability to differentiate between till units is essential in sampling and interpreting results. Multiple till units have not yet been identified in the study area, but the complex ice-flow history and the generally thick drift cover suggest that they are likely to be found with more detailed mapping.

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PRELIMINARY TRILOBITE BIOSTRATIGRAPHY OF THE COOKS BROOK FORMATION (NORTHERN HEAD GROUP), HUMBER ARM ALLOCHTHON, BAY OF ISLANDS, WESTERN NEWFOUNDLAND

W.D. Boyce, J.W. Botsford[1] and J.S. Ash
Newfoundland Mapping Section

ABSTRACT

More than one hundred and ten fossiliferous samples from the Cooks Brook formation of Humber Arm, Middle Arm and Penguin Arm have yielded previously unrecorded trilobite faunas. The collections were obtained from the 'lower Cooks Brook (Shaly) interval', Halfway Point member, Brakes Cove member, 'upper conglomerate' and 'Lower Ordovician conglomerate'. The Cooks Brook formation is a sequence dominated by deep-water carbonate gravity deposits that forms part of the Humber Arm Allochthon. It comprises approximately 350 metres of grey-black shale and lime mudstone with interstratified carbonate conglomerate.

The 'lower Cooks Brook (Shaly) Interval' contains an in situ *fauna indicative of the late Middle Cambrian* Ehmaniella *Zone. The Halfway Point member contains* in situ *and redeposited trilobite faunas indicative of the late Middle Cambrian* Ptychagnostus gibbus *and* Bolaspidella *zones (=*Ehmaniella *Zone of restricted biofacies). The Brakes Cove member contains redeposited faunas indicative of the* Cedaria *and* Crepicephalus *zones of the early Late Cambrian Dresbachian Stage. The 'upper conglomerate' contains a redeposited fauna indicative of the* Taenicephalus *Zone of the medial Late Cambrian Franconian Stage. The Bear Cove conglomerate contains redeposited faunas, which range from the* Dunderbergia *Zone of the early Late Cambrian Dresbachian Stage to the* Taenicephalus *Zone of the medial Late Cambrian Franconian Stage. Based on the youngest contained fauna, the probable age of this conglomerate is* Taenicephalus *Zone, identical to that of the 'upper conglomerate'. The Serpentine River strata contain an* in situ *fauna correlative with the* Keithia schucherti *fauna of the allochthonous Cow Head Group, and the* Saukiella junia *or* Saukiella serotina *subzone of the latest Cambrian* Saukia *Zone (Trempealeauan Stage). The 2-m-thick 'Lower Ordovician conglomerate' contains redeposited faunas that range in age from the* Rasettia magna *Subzone of the* Saukia *Zone (Trempealeauan Stage) to the* Symphysurina brevispicata *Subzone of the earliest Ordovician* Symphysurina *Zone (Canadian Series, Gasconadian Stage). Based on the youngest contained fauna, this conglomerate is of probable* Symphysurina *Zone (*Symphysurina brevispicata *Subzone) age.*

The Cooks Brook formation is probably no older than Ehmaniella *Zone (late Middle Cambrian), and the underlying Irishtown formation is no younger than* Glossopleura *Zone (medial Middle Cambrian).*

INTRODUCTION

The Cooks Brook formation is a sequence dominated by deep-water carbonate gravity deposits that forms part of the Humber Arm Allochthon of the Bay of Islands (Figures 1 and 2). It comprises approximately 350 metres of grey-black shale and lime mudstone interstratified with carbonate conglomerate (Figure 3). Originally investigated by Stevens (1965, 1970), the Cooks Brook formation was subsequently restudied by Botsford (1988a, b), where it was assigned to the base of his newly proposed Northern Head group. During the course of his fieldwork, previously unreported late Middle Cambrian to Early Ordovician trilobite faunas were collected from this deep-water carbonate slope deposit (Botsford *et al.*, 1986).

LITHOSTRATIGRAPHY

The Cooks Brook formation conformably overlies the Irishtown formation, which is the uppermost unit of the revised Curling Group of Stevens (1970). The Irishtown formation comprises grey to black shale having thick white quartzitic sandstone and conglomerate (Williams and Cawood, 1989). The Cooks Brook formation in turn is

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[1] Newfoundland and Labrador Science and Technology Advisory Council

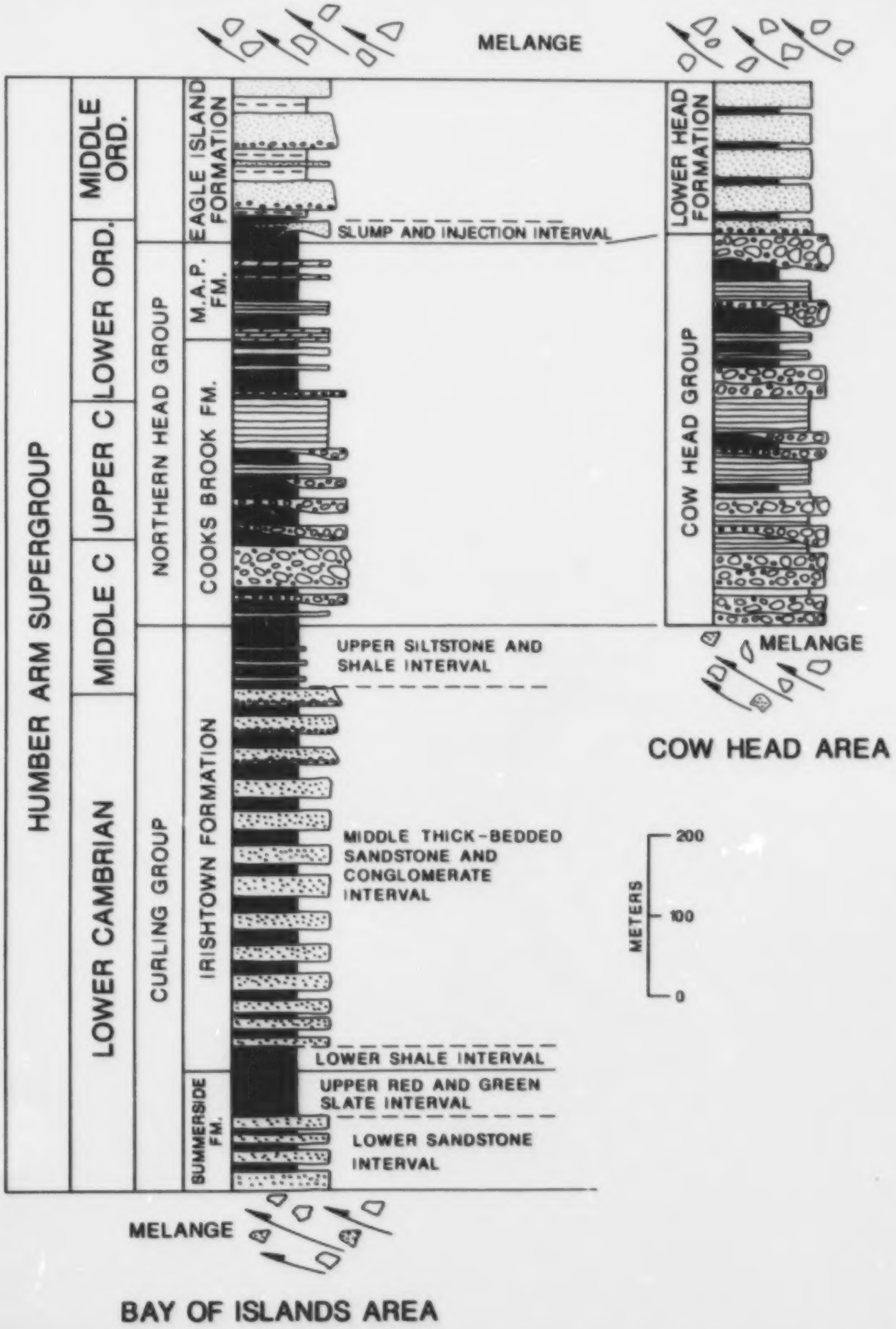
MELANGE
HUMBER ARM SUPERGROUP
MIDDLE ORD.
LOWER ORD.
UPPER C
MIDDLE C
LOWER CAMBRIAN
NORTHERN HEAD GROUP
CURLING GROUP
EAGLE ISLAND FORMATION
M.A.P. FM.
COOKS BROOK FM.
IRISHTOWN FORMATION
SUMMERSIDE FM.
SLUMP AND INJECTION INTERVAL
UPPER SILTSTONE AND SHALE INTERVAL
MIDDLE THICK-BEDDED SANDSTONE AND CONGLOMERATE INTERVAL
LOWER SHALE INTERVAL
UPPER RED AND GREEN SLATE INTERVAL
LOWER SANDSTONE INTERVAL
MELANGE
BAY OF ISLANDS AREA
LOWER HEAD FORMATION
COW HEAD GROUP
MELANGE
COW HEAD AREA
METERS
200
100
0

Figure 1. *Schematic summary of the Humber Arm Supergroup, in the Humber Arm Allochthon, illustrating the relationship of the Bay of Islands and Cow Head areas. M.A.P. is Middle Arm Point formation (from Botsford, 1988a, page 35).*

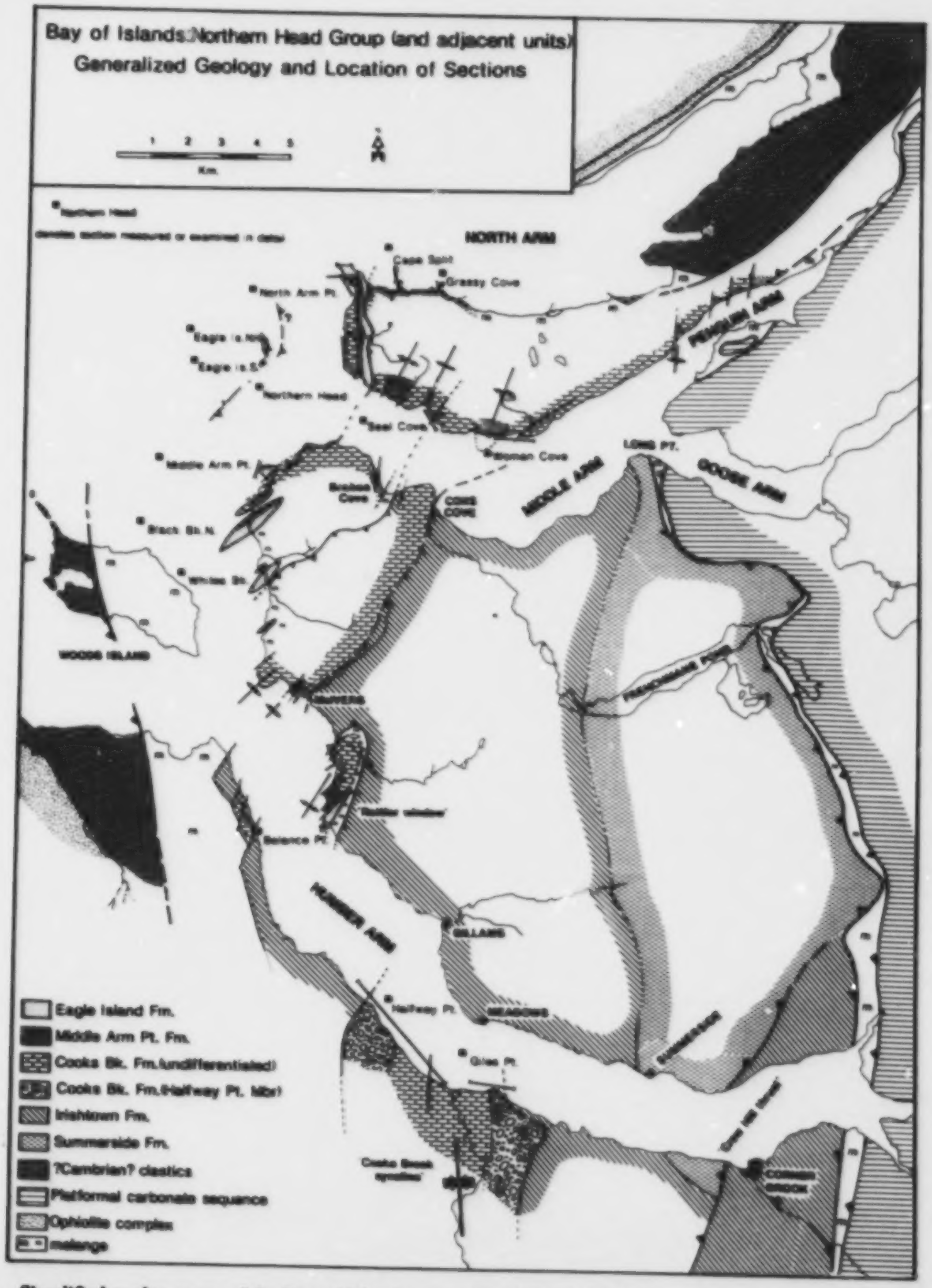
Bay of Islands: Northern Head Group (and adjacent units)
Generalized Geology and Location of Sections
1 2 3 4 5
Km.
N
Northern Head
denotes section measured or examined in detail
NORTH ARM
Cape Split
Grassy Cove
North Arm Pt.
Eagle Is. N
Eagle Is. S
Northern Head
Seal Cove
Woman Cove
Middle Arm Pt.
Brakes Cove
COXS COVE
MIDDLE ARM
LONG PT.
GOOSE ARM
PENGUIN ARM
Brach Bk. N.
Whites Bk.
WOODS ISLAND
FRENCHMANS POND
'Bottle window'
Balance Pt.
HUMBER ARM
GILLAMS
Halfway Pt.
MEADOWS
Giles Pt.
SUMMERSIDE
Cooks Brook syncline
CORNER BROOK
Eagle Island Fm.
Middle Arm Pt. Fm.
Cooks Bk. Fm.(undifferentiated)
Cooks Bk. Fm.(Halfway Pt. Mbr)
Irishtown Fm.
Summerside Fm.
?Cambrian? clastics
Platformal carbonate sequence
Ophiolite complex
melange

Figure 2. *Simplified geology map of the Bay of Islands area, showing the distribution of the Cooks Brook formation, and the locations of sampled stratigraphic sections (from Botsford, 1988a, page 42).*

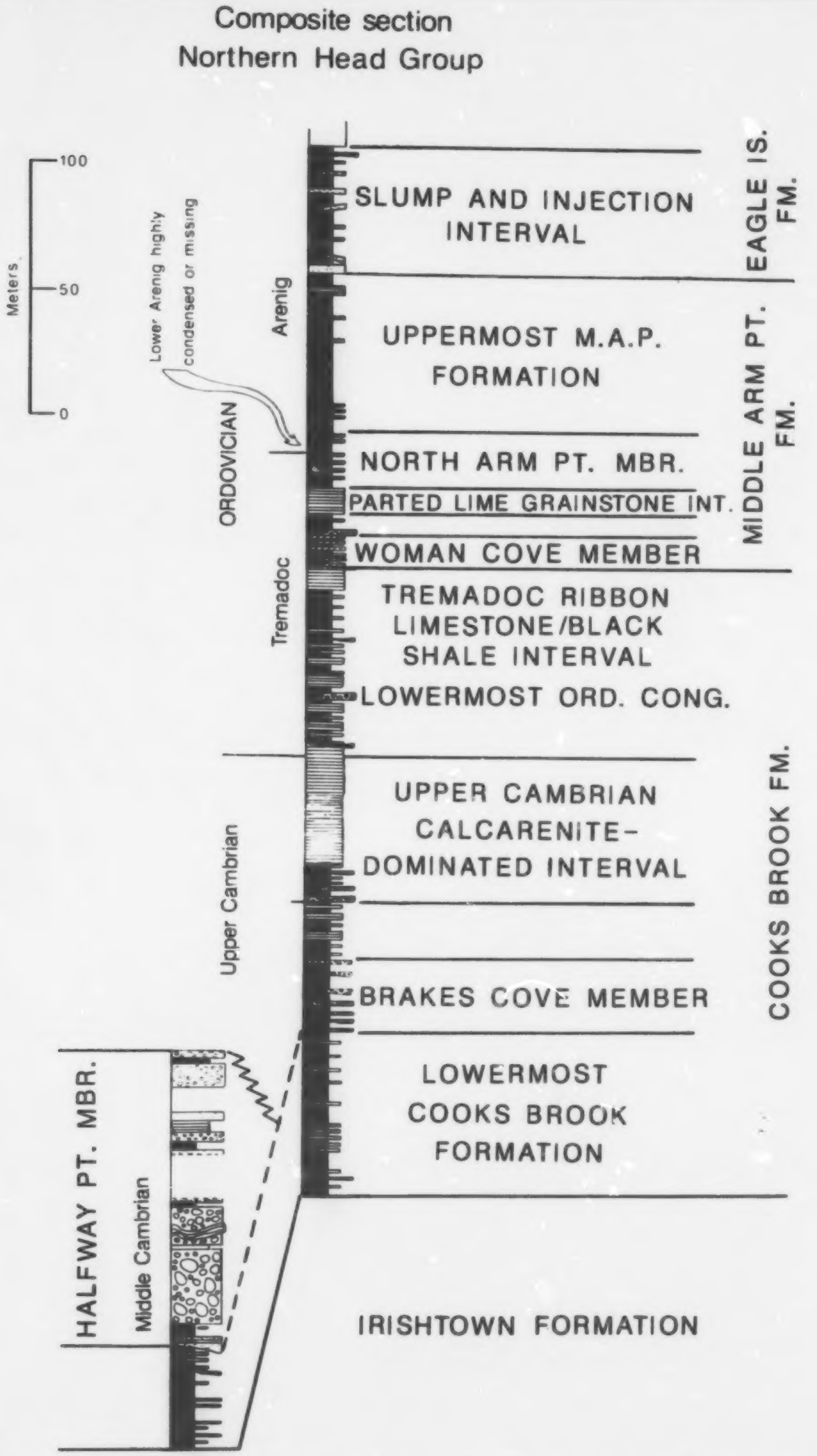
Composite section
Northern Head Group
Meters
100
50
0
Lower Arenig highly condensed or missing
Arenig
ORDOVICIAN
Tremadoc
Upper Cambrian
Middle Cambrian
HALFWAY PT. MBR.
SLUMP AND INJECTION INTERVAL
EAGLE IS. FM.
UPPERMOST M.A.P. FORMATION
NORTH ARM PT. MBR.
PARTED LIME GRAINSTONE INT.
WOMAN COVE MEMBER
MIDDLE ARM PT. FM.
TREMADOC RIBBON LIMESTONE/BLACK SHALE INTERVAL
LOWERMOST ORD. CONG.
UPPER CAMBRIAN CALCARENITE-DOMINATED INTERVAL
BRAKES COVE MEMBER
LOWERMOST COOKS BROOK FORMATION
COOKS BROOK FM.
IRISHTOWN FORMATION

Figure 3. *Composite stratigraphic section through the Northern Head group, showing the subdivisions of the Cooks Brook formation, (from Botsford, 1988a, page 56).*

conformably overlain by the Middle Arm Point formation, a sequence of thin-bedded black and green shale having associated dolomitic siltstone and minor limestone (Williams and Cawood, 1989).

PREVIOUS WORK

Schuchert and Dunbar (1934, page 90) were the first to collect trilobites from the Cooks Brook formation. They identified specimens of *Agnostus* and *Triarthrus* from limestone conglomerates exposed south of the mouth of Serpentine River.

Kindle and Whittington (1965) subsequently restudied Schuchert and Dunbar's material. Their *Triarthrus* was reidentified as *Bienvillia* sp. cf. *B. corax* (Billings, 1865). They also collected additional trilobite material from dark-coloured limestones about half a mile south of the mouth of Serpentine River. They identified the following trilobites:

Apatokephaloides? sp.
Bienvillia sp. cf. *B. corax* (Billings, 1865)
Geragnostus sp.
Leiobienvillia sp.
Loganopeltoides sp.
Richardsonella? sp.
Theodenisia sp.
Zacompsus sp.

The generic assemblage is indicative of a latest Cambrian *Saukia* Zone (Trempealeauan Stage) age. The above taxa further suggest a correlation with the *Keithia schucherti* fauna of the allochthonous Cow Head Group. Ludvigsen *et al.* (1989, page 10) report *Bienvillia corax* (Billings, 1865), as well as species of *Apatokephaloides*, *Theodenisia* and *Zacompsus* from the fauna, which they correlate with the *Saukiella junia* and *Saukiella serotina* subzones (Ludvigsen *et al.*, 1989, page 10, text-figure 12).

In Bear Cove, 10 km north of Serpentine River, Kindle and Whittington (1965) obtained more trilobites from the Cooks Brook formation, exposed on the White Rocks. From a granular limestone boulder, they reported the following trilobites:

Bathyholcus sp.
Buttsia sp.
Cheilocephalus? sp.
Dunderbergia sp.
Oligometopus sp.
Pseudagnostus sp. cf. *P. prolongus* (Hall and Whitfield, 1877)
Pterocephalia? sp.

The above trilobites are indicative of a *Dunderbergia* Zone age (late Dresbachian Stage).

From an oolitic limestone boulder at Bear Cove, Kindle and Whittington also collected *Parabolinoides*. This genus is indicative of a *Taenicephalus* Zone age (medial Franconian Stage).

At Halfway Point, Dr. R. K. Stevens (Memorial University of Newfoundland) later obtained trilobites from the Halfway Point member of the Cooks Brook formation (Botsford, 1988a, page 465; Botsford *in* Cawood *et al.*, 1988, page 56). From one boulder he collected the following trilobites:

Bathyuriscus sp.
Kootenia sp.
Peronopsis gaspensis? Rasetti, 1948
ptychopariid gen. et sp. nov?
Semisphaerocephalus sp.

From a second boulder, Dr. Stevens also obtained *?Phoidagnostus* sp. The taxa above are indicative of a late Middle Cambrian *Ptychagnostus gibbus* Zone age (A.R. Palmer, personal communication to R. K. Stevens, 1979).

BIOSTRATIGRAPHY

Between 1984 and 1986, and in 1991, the authors collected more than one hundred and ten fossiliferous samples from redeposited clasts (pebbles, cobbles, boulders), and several *in situ* horizons in the Cooks Brook formation. Three main areas were investigated: Humber Arm, Middle Arm and Penguin Arm (Figure 2). Most of the collections were obtained from the 'lowermost Cooks Brook (Shaly) interval', Halfway Point member, Brakes Cove member, 'upper conglomerate' and 'Lower Ordovician conglomerate'.

LOWERMOST COOKS BROOK (SHALY) INTERVAL

Botsford (1988a, page 61) defined the base of the Cooks Brook formation at the first appearance of bedded limestone above the shale-dominated uppermost Irishtown formation. At the Halfway Point type section, these limestones comprise 20- to 50-cm-thick granule to pebble conglomerates, commonly graded and displaying a platy fabric; laminated dolomitic grainstones locally cap the limestones or form separate beds. In the lowest 10 m of the Cooks Brook formation, the carbonates are interbedded with (commonly) laminated dark-grey to black shales, which pass upward into an interval of black-green dolomitic-banded shale. At Halfway Point, the 'lowermost Cooks Brook (shaly) interval' is about 40 m thick, and is conformably overlain by the Halfway Point member.

Approximately 30 m above the base of the Cooks Brook formation, two pygidia of *Ehmaniella cloudensis* (Howell, 1943) were collected from a limestone conglomerate lens (in matrix at the base of a small channel/scour). Previously, the only fossils from the 'lowermost Cooks Brook (Shaly) interval', were indeterminate agnostids from thin limestone beds in Middle Trout River Brook (Botsford, 1988a, page 62). *Ehmaniella cloudensis* (Howell, 1943) is the characteristic species of the late Middle Cambrian *Ehmaniella cloudensis* Zone (Knight and Boyce, 1987); it is widespread throughout western Newfoundland in the lower part of the (autochthonous) March Point Formation (Howell, 1943; Boyce, 1977; Stouge and Boyce, 1983; Knight and Boyce, 1987). The specimens are, therefore, firmly indicative of a late Middle Cambrian *Ehmaniella* Zone age (Knight and Boyce, 1987). This suggests that the Cooks Brook formation is probably no older than *Ehmaniella* Zone (late Middle Cambrian), and the underlying Irishtown formation is no younger than *Glossopleura* Zone (medial Middle Cambrian).

HALFWAY POINT MEMBER

The Halfway Point member is approximately 120 m thick; it consists predominantly of lime conglomerates separated by interbedded, parted lime grainstones and nodular to ribbon limestones and is best exposed at Halfway Point, on the south shore of Humber Arm (Botsford, 1988a, page 58).

In 1991, the Halfway Point member was again sampled for trilobites. Fifty-three collections were obtained, mainly from the basal conglomerates (*Bathyuriscus–Elrathina* Zone trilobite fauna, of Figure 4). Preliminary analysis suggests that the basal conglomerates include clasts of *Bolaspidella* Zone age, as well as *Ehmaniella* Zone and *Ptychagnostus gibbus* Zone (*Bathyuriscus–Elrathina* Zone of earlier workers). Consequently, the basal conglomerates are probably younger (*Bolaspidella* Zone) than earlier believed.

The basal conglomerates of the Halfway Point member are interrupted by several metres of shale. Below the shale, at least two distinct conglomerate layers occur; from these, seventeen fossiliferous samples were collected. Most of the collections have yet to be studied in detail. However, *?Onymagnostus/Ptychagnostus* sp. undet. occurs in a light-grey platy limestone clast (sample number (SN) 91F018C) collected from a fallen block derived from the lower conglomerate; it indicates a Middle Cambrian age. From the upper conglomerate, the following trilobite taxa were retrieved from possible matrix (SN 91F019A):

agnostid gen. et sp. undet.
Bathyuriscus sp. undet.
Modocia sp. undet.

In the Great Basin, the genus *Modocia* appears just below the base of the *Bolaspidella* Zone (Robison, 1976, page 98, text-figure 4).

Above the shale, at least three distinct conglomerate layers occur; from these, twenty-two fossil collections were obtained. From the lowest layer, the trilobites of five collections have been examined in (relative) detail. A lime

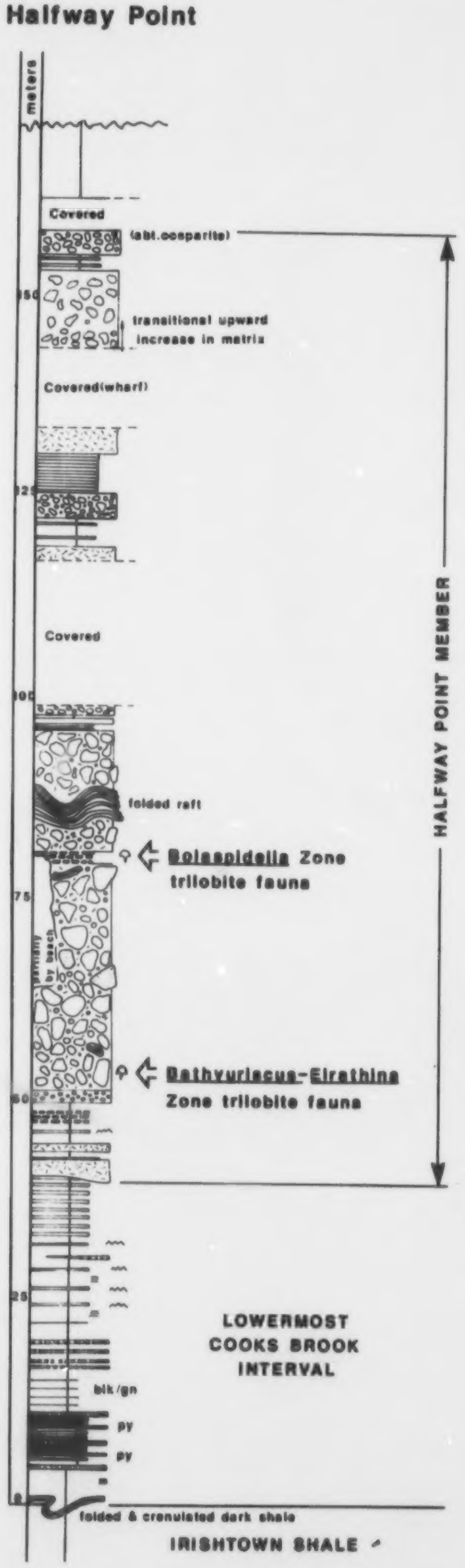
Halfway Point
meters
Covered
(abt.oosparite)
150
transitional upward increase in matrix
Covered(wharf)
125
Covered
100
folded raft
Bolaspidella Zone trilobite fauna
75
Bathyuriscus–Elrathina Zone trilobite fauna
50
HALFWAY POINT MEMBER
25
LOWERMOST COOKS BROOK INTERVAL
blk/gn
py
py
0
folded & crenulated dark shale
IRISHTOWN SHALE

Figure 4. *Stratigraphic section through the "lower Cooks Brook (Shaly) interval" and the Halfway Point member of the Cooks Brook formation (from Botsford, 1988a, page 60).*

mudstone clast (SN 91F020A) in a fallen block yielded the following:

Ehmaniella cloudensis (Howell, 1943)
Hypagnostus sp. undet.
Olenoides foveolatus Rasetti, 1948

The fortuitous occurrence of *Ehmaniella cloudensis* (Howell, 1943) and *Olenoides foveolatus* Rasetti, 1948 in the same sample allows some interesting correlations to be made. As stated above, *Ehmaniella cloudensis* (Howell, 1943) is the characteristic species of the late Middle Cambrian *Ehmaniella cloudensis* Zone in the autochthonous sequence of western Newfoundland. *Olenoides foveolatus* Rasetti, 1948 was originally described by Rasetti (1948) from Middle Cambrian boulders in the Lévis Formation of Québec; it also occurs in the *Ptychagnostus gibbus* Zone-correlative *Zacanthoides gilberti* Fauna of the allochthonous Cow Head Group in western Newfoundland (Young and Ludvigsen, 1989). There it is associated with the agnostid *Peronopsis interstricta* (White, 1874), which, according to Young and Ludvigsen (1989, page 11) is similar to *Peronopsis gaspensis* Rasetti, 1948 (previously reported from the Halfway Point member—see above). The *Ehmaniella cludensis* Zone of the autochthon and the *Zacanthoides gilberti* Fauna of the allocthon both correlate, therefore, with the *Ptychagnostus gibbus* Zone. In addition, Knight and Boyce (1987, page 361, Figure 4) report *Zacanthoides* sp. undet. from the *Olenoides longispinus* Zone in the March Point Formation of Canada Bay. Boyce *(op. cit.)* believes that this species is probably conspecific with *Zacanthoides gilberti* Young and Ludvigsen, 1989; if so, the *Olenoides longispinus* Zone probably also correlates, at least in part, with the *Zacanthoides gilberti* Fauna and the *Ptychagnostus gibbus* Zone (Table 1).

A lime grainstone cobble (SN 91F020B) from another fallen block yielded:

agnostid gen. et sp. undet.
Asaphiscus sp. cf. *A. laeviceps* (Walcott, 1884)

Asaphiscus sp. cf. *A. laeviceps* (Walcott, 1884) also occurs in the upper part of the March Point Formation in western Newfoundland (Stouge and Boyce, 1983). *Asaphiscus laeviceps* (Walcott, 1884) is a younger Middle Cambrian species than *Asaphiscus wheeleri* Meek, 1873 (Palmer, 1954). This implies a *Bolaspidella contracta* Subzone age (see below).

Another lime grainstone clast (SN 91F020D) contained the following:

Bathyuriscus sp. undet.
?Bolaspidella sp. undet.
ptychopariid? gen. et sp. undet.

?Bolaspidella sp. undet. suggests a *Bolaspidella* Zone age for this clast.

The following were obtained from a lime mudstone clast (SN 91F020F):

Pagetia sp. undet.
ptychopariid gen. et spp. undet.

Rasetti (1966, page 504) and Robison (1976, page 98, text-figure 4), state that the genus *Pagetia* ranges from the (late Early Cambrian) *Bonnia—Olenellus* Zone to the top of the (Middle Cambrian) *Oryctocephalus* Zone (*Ptychagnostus gibbus* Zone-correlative), and is entirely a pre-*Bolaspidella* Zone taxon.

Another lime mudstone clast (SN 91F020G) yielded:

Trilobita

Bathyuriscus sp. undet.
Ehmania borealis Howell, 1943

Ehmania borealis Howell, 1943 ranges through the *Glossopleura walcotti*, *Polypleuraspis*, *Olenoides longispinus* and *Ehmaniella cloudensis* zones in the autochthonous sequence of the Canada Bay area (Knight and Boyce, 1987). This widespread species most often occurs in the lower part of the March Point Formation in association with *Ehmaniella cloudensis* (Howell, 1943) (Howell, 1943; Boyce, 1977; Stouge and Boyce, 1983; Knight and Boyce, 1987); therefore, it is also probably indicative of a late Middle Cambrian *Ehmaniella* Zone age.

From the middle conglomerate layer, notable for its shaly matrix, a platy clast of buff-weathering, dark-grey dolomite mudstone—wackestone (SN 91F021C) yielded:

Asaphiscus sp. cf. *A. wheeleri* Meek, 1873
Bathyuriscus sp. undet.

Asaphiscus wheeleri Meek, 1873 occurs in the *Bathyuriscus fimbriatus* Subzone of the *Bolaspidella* Zone in Utah (Robison, 1964, page 512, text-figure 2).

Finally, from the upper conglomerate layer, three collections have merited special attention. The trilobite *Olenoides foveolatus* Rasetti, 1948 was obtained from a large, platy lime mudstone clast (SN 91F022A) near the base of the layer and below a large Irishtown Formation block. As stated above, *Olenoides foveolatus* Rasetti, 1948 also occurs in the *Ptychagnostus gibbus* Zone-correlative *Zacanthoides gilberti* Fauna of Cow Head Group.

Another rectangular (30 by 15 cm) boulder of light blue-grey weathering, dark blue-grey lime mudstone—wackestone (SN 91F022H) yielded the following:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Trilobita

Bathyuriscus fimbriatus Robison, 1964. (=*Orria* sp. of Kindle, 1982, Plate 1.2, figure 8)
ptychopariid gen. et sp. undet.

Table 1. Correlation of rock units in the Bay of Islands area. Restricted shelf trilobite zones after Lochman-Balk and Wilson (1958), Palmer (1965), Robison (1976) and Stitt (1971, 1977, 1983). Open shelf agnostoid and polymeroid zones after Öpik (1967), Palmer (1962), Robison (1976, 1988). West Newfoundland platform trilobite zonation is provisional and subject to change: Albertan portion is based on sequence in Canada Bay (Boyce *in* Knight and Boyce, 1987); Dresbachian and Franconian parts based on faunas of the Port au Port Peninsula (Boyce, 1977, unpublished); Trempealeauan and Gasconadian intervals are based on Goose Arm sequence (Boyce *in* Knight and Boyce, 1991). Cow Head trilobite zonation after Fortey *et al.* (1982), Kindle (1982), Ludvigsen *et al.* (1989) and Young and Ludvigsen (1989). The Reluctant Head Formation and the Weasel group have been placed in the same column only to save space. Dashed lines indicate uncertainty

SYSTEM	SERIES	STAGE	TRILOBITE ZONATIONS						ROCK UNITS					
			RESTRICTED SHELF		OPEN SHELF		WEST	NFLD	WEST NEWFOUNDLAND					
			POLYMEROID ZONES	POLYMEROID SUBZONES	POLYMEROID ZONES	AGNOSTOID ZONES	PLATFORM (COMPOSITE)	SLOPE (COW HEAD)	"AUTOCHTHON"		"PARAUTOCHTHON"		ALLOCHTHON	
EARLY ORDOVICIAN	CANADIAN	GASCONADIAN	*SYMPHYSURINA*	*SYMPHYSURINA WOOSTERI*			*HYSTRICURUS MILLARDENSIS*	*SYMPHYSURINA*	WATTS BIGHT FORMATION		WEASEL GROUP	UPPER CONGLOMERATE UNIT	COOKS BROOK FORMATION	
				SYMPHYSURINA BULBOSA										
				SYMPHYSURINA BREVISPICATA										LOWER ORDOVICIAN CONGLOMERATE
									BERRY HEAD FORMATION	UPPER MEMBER				
LATE CAMBRIAN			*MISSISSQUOIA*	*MISSISSQUOIA TYPICALIS*			*MISSISSQUOIA TYPICALIS*	*MISSISSQUOIA TYPICALIS*						
				MISSISSQUOIA DEPRESSA			*PLETHOPELTIS*	*PHYLACTERUS SAYLESI*						
	CROIXIAN	TREMPEALEAUAN	*SAUKIA*	*EUREKIA APOPSIS*										
				SAUKIELLA SEROTINA				*KEITHIA SCHUCHERTI*						
				SAUKIELLA JUNIA										
				RASETTIA MAGNA				*KEITHIA SUBCLAVATA*						
								ONCHONOTUS RICHARDSONI						
		FRANCONIAN	*SARATOGIA*	*DRUMASPIS TEXANA*			*TAENICEPHALUS*	*BEOTHUCKIA DUOMENTA*		MIDDLE MEMBER				
				SARATOGIA LIRAE										
			TAENICEPHALUS					*ROKSASPIS TURBINELLA*						UPPER BEAR COVE CONGLOMERATE
				PARABOLINOIDES						LOWER MEMBER				
			ELVINIA	*IRVINGELLA MAJOR*			*CAMARASPIS CONVEXUS*	ZONE 7				MIDDLE SLATE UNIT		
		DRESBACHIAN	*DUNDERBERGIA*				*DUNDERBERGIA*	ZONE 6	PETIT JARDIN FORMATION	UPPER DOLOSTONE MEMBER		LOWER CONGLOMERATE UNIT		
			PREHOUSIA				*DYTREMACEPHALUS STRICTUS*							
			DICANTHOPYGE											
			APHELASPIS			*GLYPTAGNOSTUS RETICULATUS*								BRAKES COVE MEMBER
			CREPICEPHALUS			*GLYPTAGNOSTUS STOLIDOTUS*	*CREPICEPHALUS RIVUS*	ZONE 5						
MIDDLE CAMBRIAN			*CEDARIA*			*LEJOPYGE LAEVIGATA*	*WELLERASPIS NEWFOUNDLANDENSIS*			MIDDLE LIMESTONE MEMBER				
	ALBERTAN		*ELDORADIA*		*BOLASPIDELLA*	*PTYCHAGNOSTUS PUNCTUOSUS*	*ELDORADIA DUNBARI*	ZONE 4		LOWER DOLOSTONE MEMBER	RELUCTANT HEAD FORMATION			HALFWAY POINT MEMBER
			EHMANIELLA				*EHMANIELLA CLOUDENSIS*							
								ZONE 3						
						PTYCHAGNOSTUS ATAVUS								
									MARCH POINT FORMATION					
					ORYCTOCEPHALUS	[illegible]	*OLENOIDES LONGISPINUS*							LOWER COOKS BROOK
						PTYCHAGNOSTUS PRAECURRENS								
			GLOSSOPLEURA											
						PERONOPSIS BONNERENSIS	*POLYPLEURASPIS*		PENGUIN COVE FORMATION				IRISHTOWN FORMATION	
			ALBERTELLA				*GLOSSOPLEURA WALCOTTI*	ZONE 2 + *ZACANTHOIDES GILBERTI*						
							PROLIOSTRACUS							
			PLAGIURA				*ALOKISTOCARE (AMECEPHALUS) CLEORA*							

Bathyuriscus fimbriatus Robison, 1964 ranges from the *Bathyuriscus fimbriatus* Subzone to the *Bolaspidella contracta* Subzone of the *Bolaspidella* Zone in Utah (Robison, 1964, page 512, text-figure 2). Kindle (1982; Plate 1.2, figure 8) illustrated *Orria* sp. (from Zone 3 of the Cow Head Group), which Boyce and Ash consider conspecific with *Bathyuriscus fimbriatus* Robison, 1964.

Bathyuriscus sp. cf. *B. fimbriatus* Robison, 1964 was also collected from a rounded (15 cm) cobble of light blue-grey weathering, dark blue-grey lime mudstone–wackestone (SN 91F022I). Here it should be noted that samples 91F022H and 91F022I were collected from loose blocks of Cooks Brook formation resting on the middle conglomerate layer. However, because these blocks do not have a shaly matrix, they are probably not from that layer. Consequently, their source is either the lower or upper layers. Boyce thinks they fell from the upper layer; Knight (personal communication, 1991) believes they were pushed from the lower layer by ice.

The basal conglomerates of the Halfway Point member contain faunas correlative with the restricted shelf *Ehmaniella* Zone, as well as the open shelf *Ptychagnostus gibbus* (agnostid) Zone and *Bolaspidella* (polymeroid) Zone. Based on their youngest contained faunas, they are of probable *Bolaspidella* Zone (*Bathyuriscus fimbriatus* Subzone) age. Consequently, they are slightly younger than earlier believed.

Trilobites were also collected from three *in situ* horizons near the top of the lower half Halfway Point member (*'Bolaspidella* Zone trilobite fauna' of Figure 4). The thin-bedded, nodular dolomitic lime mudstone yielded the following composite fauna:

Bathyuriscus elegans (Walcott, 1916)
Hemirhodon sp. cf. *H. amplipyge* Robison, 1964
Modocia nuchaspina Robison, 1964

In Utah, *Bathyuriscus elegans* (Walcott, 1916) and *Hemirhodon amplipyge* Robison, 1964 occur in the upper half of the *Bolaspidella contracta* Subzone of the *Bolaspidella* Zone; *Modocia nuchaspina* Robison, 1964 is restricted to the base of the *Bolaspidella contracta* Subzone (Robison, 1964, text-figure 2). The *in situ* fauna of the Halfway Point member, therefore, correlates with that of the *Bolaspidella contracta* Subzone of the *Bolaspidella* Zone, as well as that of Zone 4 of the Cow Head Group (Kindle, 1982); it probably also correlates with that of the upper part of the parautochthonous Reluctant Head Formation (Knight and Boyce, 1991; Boyce *et al., this volume*).

The above fauna was the highest obtained from the Halfway Point member. Consequently, the uppermost strata of the unit may be as young as *Cedaria* Zone.

BRAKES COVE MEMBER

The Brakes Cove member is 12 to 15 m thick; it consists of 0.5- to 2.0-m-thick units of pebble to cobble conglomerate interbedded on a 3 m scale with nodular to ribbon limestone (Botsford, 1988a, page 62).

Fossiliferous boulders were collected from the Brakes Cove member at Brakes Cove, Northern Head and Woman Cove Head in Middle Arm, and at Allans Brook in Penguin Arm.

Cedaria Zone Faunas

For the purposes of this preliminary report, fossil collections characterized by specimens of *Cedaria* **only** are correlated with the *Cedaria* Zone of the early Late Cambrian Dresbachian Stage.

At Northern Head, the following composite fauna was obtained from seven boulders (SN 85F092, 85F094, 85F095, 85F099, 85F101-103):

Brachiopoda–Inarticulata

Gen. et sp. undet.

Trilobita

Balderia sp. cf. *B. aspera* Robison, 1988
?Bonneterrina sp. undet.
Cedaria sp. undet.
Holcacephalus tenerus (Walcott, 1916)
Hypagnostus sp. undet.
Kingstonia sp. undet.
Kormagnostus seclusus (Walcott, 1884)
Tricrepicephalus sp. undet.

At Woman Cove Head, the following composite fauna was obtained from eight boulders (SN 85F106, 85F107, 85F109-85F111, 85F113-85F115):

Trilobita

?Bolaspidella sp. of Kindle (1982)
Cedaria sp. undet.
?Cedaria sp. undet.
Exigua quebecensis (Rasetti, 1946)
Hemirhodon sp. cf. *H.* sp. 1 of Robison (1988)
Hypagnostus sp. undet.
Kormagnostus seclusus (Walcott, 1884)
Onchonotopsis pergibba Rasetti, 1946
Peronopsis tenuis (Illing, 1916)
Syspacheilus sp. undet.
Tricrepicephalus sp. undet.
Welleraspis newfoundlandensis Lochman, 1938
Welleraspis sp. cf. *W.* sp. of Palmer *in* Palmer and Peel, 1981, Plate 6, figures 5-7)

Balderia aspera Robison, 1988, *Cedaria prolifica* Walcott, 1924, *Exigua quebecensis* (Rasetti, 1946), *Hemirhodon* sp. 1 of Robison (1988), *Kormagnostus seclusus* (Walcott, 1884), *Onchonotopsis pergibba* Rasetti, 1946, *Peronopsis tenuis*

(Illing, 1916) and *Welleraspis newfoundlandensis* Lochman, 1938 all occur in *Cedaria* Zone-correlative strata of the Holm Dal Formation in north Greenland (Robison, 1988, pages 26-29).

?Bolaspidella sp. of Kindle (1982) occurs in Boulder 468 of Kindle's (Cow Head Group) Zone 4, which he correlated with the uppermost part of the *Bolaspidella* Zone. *Cedaria prolifica* Walcott, 1924 ranges from the uppermost part of the *Lejopyge laevigata* Zone (*Cedaria* Zone-correlative) to the *Glyptagnostus stolidotus* Zone (*Crepicephalus* Zone-correlative) (Robison, 1988, page 59). *Exigua quebecensis* (Rasetti, 1946), *Holcacephalus* sp. cf. *H. tenerus* (Walcott, 1916) and *Onchonotopsis pergibba* Rasetti, 1946 also occur in Upper Cambrian boulders in the Lévis Formation of Québec (Rasetti, 1946b). *Holcacephalus tenerus* (Walcott, 1916) occurs in both the *Cedaria* Zone and the *Crepicephalus* Zone of Montana (Lochman and Duncan, 1944, page 136).

Robison (1988, page 45) stated that *Kormagnostus seclusus* (Walcott, 1884) is an exceptionally variable and widespread species, which ranges from the lower *Lejopyge laevigata* Zone (*Bolaspidella* Zone-correlative) to the *Glyptagnostus stolidotus* Zone (*Crepicephalus* Zone-correlative).

Peronopsis tenuis (Illing, 1916) ranges from the *Ptychagnostus gibbus* Zone to the *Lejopyge laevigata* Zone (Robison, 1988, page 48). However, the pygidium of *P. tenuis* from the Brakes Cove member displays the secondary median node characteristic of *P. tenuis* of the *Lejopyge laevigata* Zone (latest *Bolaspidella* Zone- to *Cedaria* Zone-correlative).

Welleraspis newfoundlandensis Lochman, 1938 is also found in *Cedaria* Zone-correlative beds in the Petit Jardin Formation of the Port au Port Peninsula, western Newfoundland (Lochman, 1938), and in the middle limestone member of the Petit Jardin Formation in Goose Arm, western Newfoundland (Knight and Boyce, 1991). *Welleraspis* sp. of Palmer *in* Palmer and Peel (1981) occurs in *Cedaria* Zone-correlative strata of the Cass Fjord Formation in north Greenland (Palmer and Peel, 1981, page 17).

Cedaria Zone to *Crepicephalus* Zone Faunas

Faunas characterized by specimens of *Deiracephalus*, *Holcacephalus tenerus* (Walcott, 1916), *Kingstonia*, *Kormagnostus seclusus* (Walcott, 1884), *Pemphigaspis* and *Tricrepicephalus* (and lacking either *Cedaria* and *Crepicephalus*) are correlated with a late *Cedaria* to *Crepicephalus* zonal age because these taxa range between the two zones.

At the type section in Brakes Cove, Botsford (1988a) obtained the following fauna from 6 boulders (SN 86F088, 86F091, 86F093, 86F094, 86F096, 86F097):

Trilobita

Deiracephalus sp. undet.
Densonella sp. undet.
Gen. et sp. undet.
Holcacephalus tenerus (Walcott, 1916)
Kingstonia sp. undet.
Kormagnostus seclusus (Walcott, 1884)
?Meteoraspis sp. undet.
Tricrepicephalus sp. undet.

Another boulder (SN 86F099) yielded the following:

Trilobita

Catillicephala sp. undet.
Cedaria sp. cf. *C. brevifrons* Palmer, 1962
Cedaria prolifica Walcott, 1924
Modocia sp. cf. *M. nevadensis* Palmer, 1954
Pemphigaspis bullata Hall, 1863

As stated above, *Cedaria prolifica* Walcott, 1924 ranges from the *Cedaria* Zone to the *Crepicephalus* Zone. *Modocia nevadensis* Palmer, 1954 occurs with *Asaphiscus laeviceps* (Walcott, 1884) in *Bolaspidella* Zone-correlative strata in Nevada (Palmer, 1954). *Pemphigaspis bullata* Hall, 1863 was also formerly known as *Hallaspis matutina* (Hall, 1863)—see Palmer (1951). It occurs in the early Dresbachian *Cedaria* and *Crepicephalus* zones of Minnesota and Wisconsin (Raasch and Lochman, 1943, page 231). *Pemphigaspis bullata* Hall, 1863 also occurs in *Crepicephalus* Zone-correlative strata of the Warrior Formation in Pennsylvania (Tasch, 1951) and the Bonneterre Dolomite of Missouri (Lochman, 1968).

At Seal Cove, one boulder (SN 85F230) yielded the following:

Trilobita

?Bynumina sp.
Deiracephalus sp. undet.
Gen. et sp. undet.
Kingstonia sp. undet.
?Sigmocheilus sp.

At Allans Brook, Penguin Arm, the following was collected from two boulders (SN 86F082, 86F084):

Trilobita

Kingstonia sp. undet.

Crepicephalus Zone Faunas

For the purposes of this preliminary report, any fossil collections containing specimens of *Crepicephalus* are correlated with the *Crepicephalus* Zone of the early Late Cambrian Dresbachian Stage.

At the type section in Brakes Cove, the following composite fauna was obtained from three boulders (SN 86F090, 86F092, 86F095):

Trilobita

Crepicephalus sp. undet.
Holcacephalus tenerus (Walcott, 1916)

Hypagnostus sp. undet.
Kormagnostus seclusus (Walcott, 1884)
Meteoraspis sp. undet.
Pseudagnostus sp. cf. *P. prolongus* (Hall and Whitfield, 1877)
Tricrepicephalus sp. undet.

At Northern Head, the following composite fauna was obtained from twelve boulders (SN 85F089, 85F090, 85F093, 85F096-85F098, 85F100, 85F104, 85F226-85F229):

Trilobita

Blainiopsis sp. cf. *B. holtedahli* Poulsen, 1946
Blountia/Coosella sp. undet.
Bynumina sp. undet.
Cedaria sp. undet.
?Cedaria sp. undet.
Crepicephalus sp. undet.
Coosella/Meteoraspis sp. undet.
?Coosella sp. undet.
?Deiracephalus sp. undet.
?Eldoradia sp. undet.
Gen. et sp. undet.
Holcacephalus tenerus (Walcott, 1916)
Hypagnostus sp. undet.
?Hypagnostus sp. undet.
Kingstonia sp. undet.
Kormagnostus seclusus (Walcott, 1884)
?Modocia sp. undet.
Onchonotopsis sp. cf. *O. occidentalis* Palmer, 1968
Ptychagnostus/Trilobagnostus sp. undet.
Tricrepicephalus sp. undet.
?Welleraspis sp. undet.

In Alaska, *Onchonotopsis occidentalis* Palmer, 1968 occurs with *Lejopyge calva* Robison, 1964 (Palmer, 1968, page B92), a species indicative of a *Lejopyge laevigata* Zone age (latest *Bolaspidella* Zone–to *Cedaria* Zone-correlative–see Table 1).

At Allans Brook, Penguin Arm, Botsford obtained the following from one boulder (SN 86F083):

Trilobita

Bienella sp. cf. *B. problematica* Lochman *in* Lochman and Duncan (1944)

Bienella problematica Lochman *in* Lochman and Duncan, 1944 occurs in the *Crepicephalus* Zone of Montana (Lochman and Duncan, 1944, page 65).

To summarize, the Brakes Cove member contains faunas indicative of the *Cedaria* and *Crepicephalus* zones (early Dresbachian Stage), which are comparable to those of Zone 5 in the allochthonous Cow Head Group (Kindle, 1982). They also correlate with the faunas of the middle limestone and upper dolostone members of the autochthonous Petit Jardin Formation of Goose Arm (Knight and Boyce, 1991) and those of the lower conglomerate unit of the parautochthonous Weasel group (Boyce *et al.*, *this volume*).

UPPER CONGLOMERATE

The 'upper conglomerate' is best exposed at Seal Cove. It is a 30-m-thick-interval dominated by 0.2- to 0.8 m-thick lensoid pebble conglomerate units; these are interbedded with abundant green shale and 1 to 2 m packages of nodular to ribbon limestone (Botsford, 1988a, page 67).

One fossiliferous boulder collected by Botsford (1988a) yielded the following (SN 85F231):

Trilobita

?Dellea sp. undet.
Orygmaspis (Parabolinoides) hebe (Frederickson, 1949)

The above trilobites are indicative of a *Taenicephalus* Zone age (medial Franconian Stage).

LOWER ORDOVICIAN CONGLOMERATE

The 'Lower Ordovician conglomerate' is best exposed immediately east of the tip of Northern Head proper. It is a 2-m-thick unit of ungraded pebble to boulder conglomerate with planar boundaries. The conglomerate occurs within a sequence of parted to ribbon lime grainstones (Botsford, 1988a, page 72).

From one boulder (SN 85F120) at Northern Head, Dr. P.A. Cawood (Memorial University of Newfoundland) obtained for us a pygidium of *Platydiamesus inornatus* (Raymond, 1924). This species is indicative of a *Saukia* Zone age; it is known from the Gorge Formation of Vermont, the Lévis Formation of Québec, and the Shallow Bay Formation (Cow Head Group) of western Newfoundland (Ludvigsen *et al.*, 1989, page 38). In Québec and Vermont (Raymond, 1924, 1937; Rasetti, 1944, 1945, 1946a, 1963; Clark and Shaw, 1968), it is associated with species that also occur in the *Rasettia magna*, *Saukiella junia*, *Saukiella serotina* and *Eurekia apopsis* subzones of the *Saukia* Zone in Oklahoma (Stitt, 1977). In western Newfoundland, *Platydiamesus inornatus* (Raymond, 1924) occurs in the *Keithia subclavata* and the *Keithia schucherti* faunas of Ludvigsen *et al.* (1989, pages 9-10).

Four other boulders at Northern Head (SN 85F118, 85F119, 85F121, 85F122) yielded the following composite fauna:

Brachiopoda–Articulata

Apheorthis sp. undet.
?Nanorthis sp. undet.

Trilobita

Ptychopleurites brevifrons (Kobayashi, 1936)
Symphysurina bubops Winston and Nicholls, 1967

Ptychopleurites brevifrons (Kobayashi, 1936) occurs in the *Mississquoia depressa* Subzone of the *Mississquoia* Zone in Oklahoma (Stitt, 1977; Miller *et al.*, 1982, text-figure 4). It also occurs in the *Mississquoia depressa* Subzone of the *Parabolinella* Zone in the District of Mackenzie (Ludvigsen, 1982a, b).

In Oklahoma, central Texas and Utah, *Symphysurina bubops* Winston and Nicholls, 1967 is restricted to the *Symphysurina brevispicata* Subzone of the *Symphysurina Zone* (Stitt, 1977; Miller *et al.*, 1982, text-figures 4, 7, 8).

Based on the youngest contained fauna, the age of the 'Lower Ordovician conglomerate' is most likely *Symphysurina* Zone (*Symphysurina brevispicata* Subzone)–Canadian Series, Gasconadian Stage. Consequently, it correlates with the uppermost part of the autochthonous Berry Head Formation (Port au Port Group), and at least the basal portion of the Watts Bight Formation (St. George Group) exposed along Goose Arm (Knight and Boyce, 1991, page 146, Figure 3). This fauna probably also correlates with those of the *Symphysurina* conglomerates of the Cow Head Group, exposed at Broom Point (Fortey *et al.*, 1982).

SUMMARY

Trilobite species indicate that the Cooks Brook formation ranges in age from the late Middle Cambrian *Ptychagnostus gibbus* Zone to the Early Ordovician *Symphysurina* Zone. The Cooks Brook formation, therefore, correlates with Beds 0 to 8 of the allochthonous Cow Head Group (James and Stevens, 1986). It also broadly correlates with the autochthonous March Point, Petit Jardin and Berry Head formations (Port Au Port Group) and the Watts Bight Formation (St. George Group). The Halfway Point member contains *in situ* and redeposited trilobite faunas indicative of the late Middle Cambrian *Ptychagnostus gibbus* and *Bolaspidella* zones (=*Ehmaniella* Zone of restricted biofacies). The Brakes Cove member contains redeposited faunas indicative of the early Late Cambrian *Cedaria* and *Crepicephalus* zones. The 'Upper Conglomerate' contains a redeposited fauna indicative of the medial Late Cambrian *Taenicephalus* Zone. The Bear Cove conglomerate of Kindle and Whittington (1965) contains redeposited faunas that range from the early Late Cambrian *Dunderbergia* Zone to the medial Late Cambrian *Taenicephalus* Zone. Based on the youngest contained fauna, the probable age of this conglomerate is *Taenicephalus* Zone, identical to that of the 'upper conglomerate'. The Serpentine River strata of Kindle and Whittington (1965) contain an *in situ* fauna indicative of the *Saukiella junia* or *Saukiella serotina* Subzone of the latest Cambrian *Saukia* Zone. The 2-m-thick 'Lower Ordovician' conglomerate contains redeposited faunas that range in age from the *Rasettia magna* Subzone of the latest Cambrian *Saukia* Zone to the *Symphysurina brevispicata* Subzone of the earliest Ordovician *Symphysurina* Zone. Based on the youngest contained fauna, this conglomerate is probably *Symphysurina* Zone (*Symphysurina brevispicata* Subzone) in age.

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THE WEASEL GROUP, GOOSE ARM AREA, WESTERN NEWFOUNDLAND: LITHOSTRATIGRAPHY, BIOSTRATIGRAPHY, CORRELATION AND IMPLICATIONS

W.D. Boyce, I. Knight and J.S. Ash
Newfoundland Mapping Section

ABSTRACT

The Weasel group is a predominantly upright but allochthonous 300-m-thick sequence of suspension-deposited and resedimented carbonates and slates, which occurs immediately below the base of the Humber Arm Allochthon. Rocks are dominated by thick carbonate conglomerates and by fine-grained slightly calcareous siliciclastics (argillites and shales) that are now slates. Also present are lesser nodular and ribbon-bedded dolomitic and shaly limestone and oolitic grainstone. The succession is divided into four parts: a basal, poorly exposed interval of ribbon limestone and slate associated with minor conglomerate and a few possible slump beds; a lower carbonate conglomerate unit of Dresbachian age; a middle slate unit associated with minor fine-grained and rare conglomeratic and grainy carbonate of Franconian age; and an upper carbonate conglomerate unit of Late Cambrian to Ordovician age. The Late Cambrian conglomerates are dolomitized and rich in quartz sand.

The slates, ribbon limestones and conglomerates in the lower part of the formation are lithologically very like the older and also allochthonous Reluctant Head Formation of the Old Mans Pond area. However, they bear no true resemblance to deep-water coeval deposits of the Cow Head Group and the Cooks Brook formation of the Humber Arm Allochthon.

Approximately forty fossiliferous samples from the lower 100 m of the Weasel group have yielded three distinct and previously unrecorded trilobite and inarticulate brachiopod faunas. These redeposited and in-situ faunas are indicative of the Late Cambrian Crepicephalus, Dunderbergia *and* Elvinia *zones and show that the base of the formation is younger (early Late Cambrian,* Crepicephalus *Zone) than the top of the Reluctant Head Formation, which is of late Middle Cambrian* Bolaspidella *Zone age.*

The carbonate deposits of the Weasel group are interpreted as resedimented gravity-flow and storm-generated suspension deposits laid down on a distally steepened ramp and toe of slope setting during the Late Cambrian and Ordovician. Most of the clasts in the conglomerates are slope-derived carbonate muds. However, some skeletal and intraclastic grainstone–packstone clasts that compare to rock types logged at the top of the Reluctant Head Formation immediately below true shelf carbonates indicate a shallow water, carbonate sand shoal complex higher on the ramp.

The Weasel group and Reluctant Head Formation probably formed part of a common margin that is believed to have evolved from a prograding ramp during the Middle to early Late Cambrian to a distally steepened ramp and perhaps a true accretionary rimmed shelf during the latest Cambrian and throughout the Ordovician. The shale-dominated interval with Elvinia *fauna suggests that there was a marked sea-level rise during the Franconian Stage, followed by renewed shelf progradation in the Trempealeauan Stage. The pervasive dolomitization and quartz-sand content of the conglomerates at this time fits with geological characteristics of coeval rocks in the Cow Head Group and shelf carbonates preserved in western Newfoundland and is possibly linked to this renewed accretion.*

INTRODUCTION

At the southern edge of the Lomond (NTS 12H/5) map area (Figure 1), Nyman *et al.* (1984, pages 158-159, Figure 23.1) and later, Williams and Cawood (1989) documented a previously unrecognized sequence of thin-bedded grey limestone and buff shale, dark-grey shale and limestone (with local limestone breccia), and limestone conglomerate having a sandy limestone matrix. This sequence, informally named the Weasel group, was interpreted to occupy a separate structural slice – the 'Weasel slice' – at the base of the Humber Arm Allochthon (Figure 2) (Nyman *et al.*, 1984, page 161). The group is best exposed along a number of

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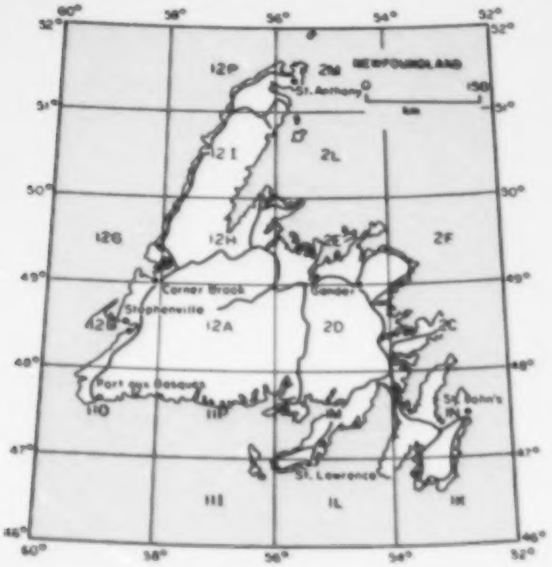


Figure 1. *Location of study area in Newfoundland.*

logging roads that occur just north of Goose Arm Brook North. Beds for the most part strike west and dip moderately to the north, but in the southern part of the slice, the group is folded and faulted. The 'Weasel slice' structurally overlies Ordovician flysch of the Goose Tickle group including limestone conglomerate of the Daniel's Harbour member (Stenzel *et al.*, 1990). These rocks form part of a thrust slice of lower Paleozoic shelf carbonates. The 'Weasel slice' is overlain structurally by dark grey, shaly melange at the base of the Humber Arm Allochthon. Based upon the sedimentary rock characteristics of the Weasel group described below and on the structural position of the 'Weasel slice', the latter is here considered to be part of the structurally disrupted and foreshortened autochthon rather than the Humber Arm Allochthon.

Botsford (1988, page 90) was the first to collect macrofossils from the Weasel group; he reported brachiopods of possible Ordovician age. During 1991, the authors discovered additional macrofossils in the unit. Boyce and Ash obtained forty-one collections from the section; the stratigraphic locations of each were subsequently determined by Knight, who measured a detailed lithostratigraphic section (Figure 3). Conodonts of Middle Ordovician age have been recovered from the conglomerates at the top of the formation (P. Cawood and F. O'Brien, personal communications, 1991).

LITHOSTRATIGRAPHY

The stratigraphy of the Weasel group is incompletely understood because of covered intervals and structural complications affecting the road section. The western end of the section in particular is folded, and it is difficult to be sure of the younging direction of beds. It is likely, however, that at least some of this part of the section lies stratigraphically below the section illustrated in Figure 3. Fossils from some conglomerates in this interval indicate that they correlate with the conglomerates at the base of the measured section.

The poorly exposed and deformed basal unit of dominantly ribbon limestone and phyllite is interbedded with some slump beds and a few limestone conglomerates, 0.05 to 1.0 m thick. The ribbon limestones (Plate 1) are planar thin-bedded and laminated, dark blue-grey with yellow- to orange-weathering, dolomitic, thin beds and phyllitic interbeds and partings. These rock types also make up the slump beds. The conglomerate consists of platy- and chip-shaped clasts set in a phyllitic carbonate matrix. However, one conglomerate contained well-rounded, medium- to coarse-grained quartz sand. The ribbon limestone and associated rock types of this interval compare well with the rocks of the Reluctant Head Formation (Knight and Boyce, 1991).

Overlying this interval of unknown thickness, the succession (Figure 3) consists of a 44 m section dominated by thick limestone conglomerate and interbeds of slate, nodular dolomitic argillite and rare ribbon limestone upward (lower conglomerate unit). Clasts (Plate 2) in these conglomerates are dominantly cm-scale plates of lime mudstone, but there are also some larger, rounded clasts up to 40 by 10 cm in size of skeletal–intraclastic grainstone, intraclastic grainstone, oolitic grainstone and intraclastic dolostone. Trilobites are present in many of these clasts. The conglomerates are dominantly clast-supported within a minor matrix comprising dolomitic fine-grained limestone. Conglomerates dominate sections up to 18 m in thickness, but the presence of 1 to 8 cm interbeds of laminated lime mudstone, locally cut out by erosion, indicates that the conglomerates are amalgamated, with beds ranging from 1 to 5 m in thickness. Fine-grained interbeds are generally poorly exposed but comprise 1 to 11 m of dark-grey slates, dolomitic argillite with calcareous nodules, and thin-bedded calcareous and dolomitic phyllite.

Olive-brown-weathering, dark-grey calcareous argillite and grey slate dominate the next 50 m (middle slate unit). These rocks may also underlie a 50-m covered interval that separates the fine-grained siliciclastics from the next overlying conglomerate unit. The slates and argillites are characterized by thin bedding and some calcareous and dolomitic laminae, and locally have a slurry-like appearance associated with small folds, suggesting slumping. The thin beds in the argillite are arranged in limestone–shale and siltstone–shale couplets suggesting, at least in the latter, that they are graded beds. Locally, the argillites are interbedded with centimetre- to decametre-scale intervals of limestone nodules and dolomitic ribbon limestone (Plate 3). Micro-load casts occur at the base of some of the ribbon limestone beds. Rare lenses of small pebble, limestone conglomerate, 3 cm thick and 10 cm wide, also occur.

At the top of this interval, there are two couplets, 1.2 and 3.3 m thick respectively, of oolitic grainstone, which are

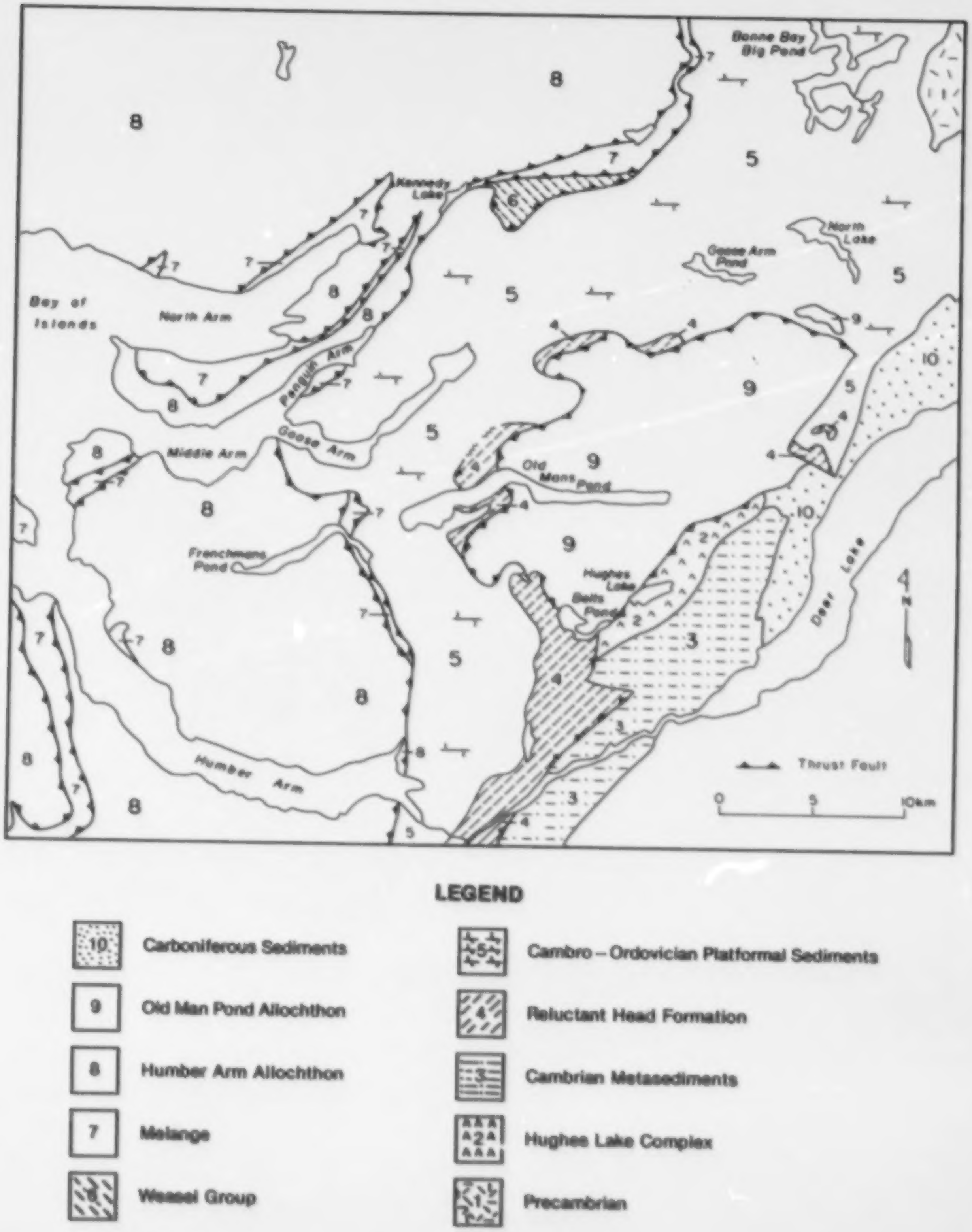
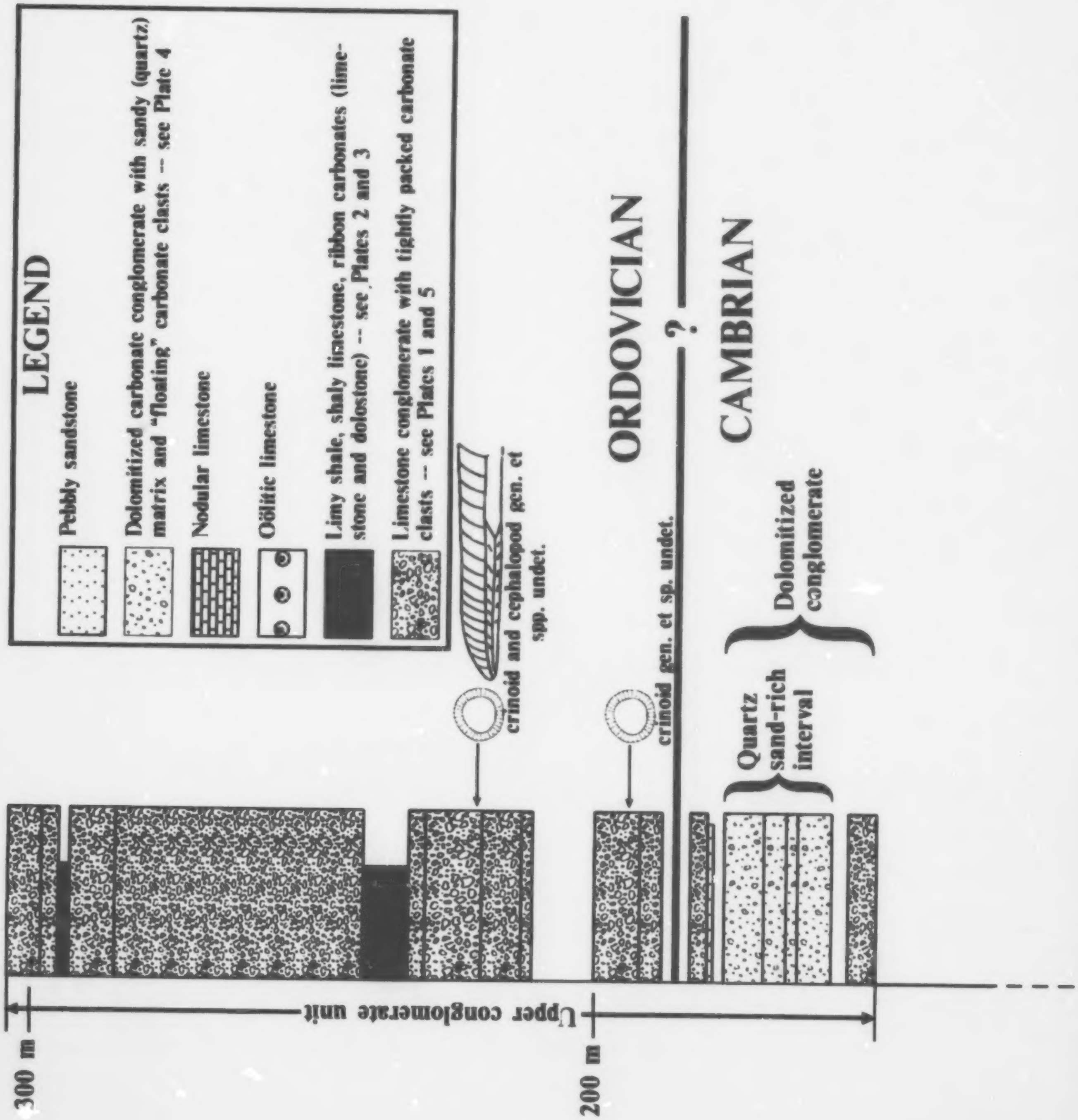

Bonne Bay Big Pond
Kennedy Lake
North Lake
Goose Arm Pond
Bay of Islands
North Arm
Penguin Arm
Middle Arm
Goose Arm
Old Mans Pond
Frenchmans Pond
Hughes Lake
Betts Pond
Deer Lake
Humber Arm
N
Thrust Fault
0
5
10km
LEGEND
10 Carboniferous Sediments
9 Old Man Pond Allochthon
8 Humber Arm Allochthon
7 Melange
6 Weasel Group
5 Cambro – Ordovician Platformal Sediments
4 Reluctant Head Formation
3 Cambrian Metasediments
2 Hughes Lake Complex
1 Precambrian

Figure 2. *Simplified geology of the Goose Arm–Humber Arm area showing the location of the study area (after Williams and Cawood, 1989).*


LEGEND
Pebbly sandstone
Dolomitized carbonate conglomerate with sandy (quartz) matrix and "floating" carbonate clasts -- see Plate 4
Nodular limestone
Oölitic limestone
Limy shale, shaly limestone, ribbon carbonates (limestone and dolostone) -- see Plates 2 and 3
Limestone conglomerate with tightly packed carbonate clasts -- see Plates 1 and 5
crinoid and cephalopod gen. et spp. undet.
crinoid gen. et sp. undet.
ORDOVICIAN
?
CAMBRIAN
Quartz sand-rich interval
Dolomitized conglomerate
Upper conglomerate unit
300 m
200 m

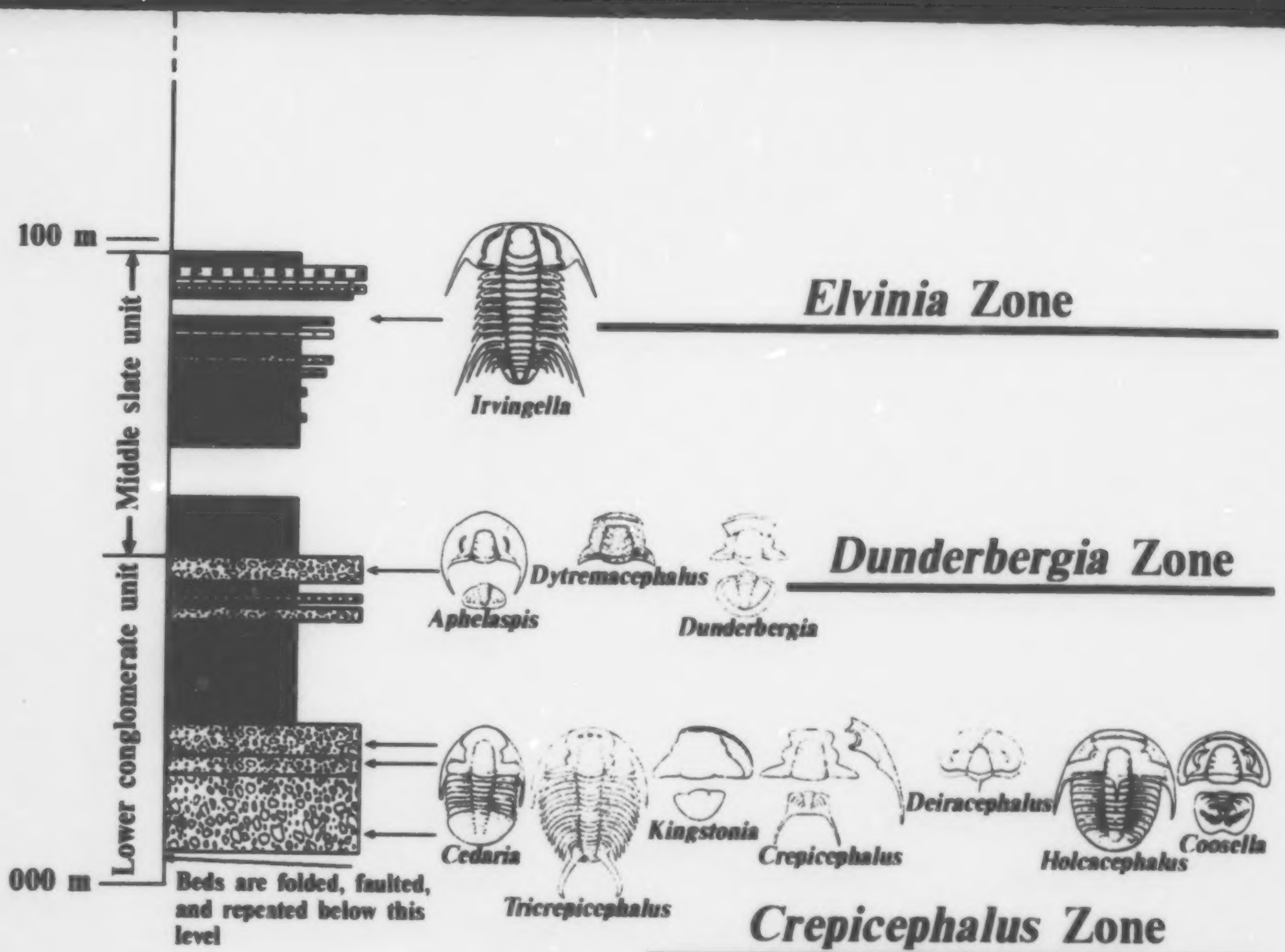

100 m
Middle slate unit
Lower conglomerate unit
000 m
Beds are folded, faulted, and repeated below this level
Irvingella
Elvinia Zone
Aphelaspis
Dytremacephalus
Dunderbergia
Dunderbergia Zone
Cedaria
Tricrepicephalus
Kingstonia
Crepicephalus
Deiracephalus
Holcacephalus
Coosella
Crepicephalus Zone

Figure 3. *Stratigraphy section through the Weasel group showing locations of fossil collections, and characteristic taxa. Trilobite reconstructions are after Moore (1959), except for* Irvingella, *which is after Rushton (1967); cephalopod and crinoid cartoons are after Flower (1976) and Sprinkle* in *Sprinkle and Kier (1987).*

Plate 1. *Dolomitic ribbon limestones of the Weasel group; these are comparable to similar lithofacies in the Reluctant Head Formation.*

Plate 2. *Limestone conglomerate near the base of the Weasel group, showing tightly packed, platy clasts of lime mudstone; lamination is visible locally in the clasts.*

Plate 3. *Planar thin-bedded, finely laminated, dark-grey calcareous argillite and shaly, dolomitic limestone below* Irvingella *locality.*

overlain by intraclastic, limestone conglomerates that are separated by a 40 cm bed of argillite. The oolitic grainstone beds are 60 cm thick and consist of fine- to medium-grained ooids, mixed in the upper bed with intraclastic lime sand-grains. The overlying conglomerates, which are 0.6 and 2.7 m thick respectively, are matrix-supported grading up into clast-supported in the higher bed. Oolitic and intraclastic sand occurs in the matrix. Clasts are dominantly platy lime mudstone having some oolitic and intraclastic grainstone clasts.

The top 150 m of the group is dominated by a clast-supported carbonate conglomerate (upper conglomerate unit). This unit is subdivided into three parts. In the first 33 m, the conglomerates are all dolomitized and include a middle interval, 18 m thick, rich in quartz sand. These dolomitized conglomerates pass upward into a succession of limestone conglomerates. At 83 m from the base of the unit, 8 m of shaly argillite and calcareous argillite interrupts the succession of limestone conglomerate.

The dolomitized conglomerate comprises buff-weathering, pale grey to grey, fine-grained dolomite and is arranged in unsorted and normally graded beds, 1 to 3 m thick. Clasts in the conglomerate are generally flat plates of buff-weathering, fine-grained dolomite to dark grey to black, grainy dolomite. Many clasts display lamination. In the middle of the section, the conglomerate matrix consists of abundant quartz sand (Plate 4), which also forms beds up to 20 cm thick of massive and crossbedded quartz arenite that gradationally cap some conglomerate beds. The quartz sand is well rounded, either fine grained or medium to coarse grained in different beds and diminishes in quantity upward in the section.

Conglomerates in the remaining part of the upper section are all limestone. They consist of metre-thick, welded beds that are poorly sorted, clast-supported, ungraded to normally graded (Plate 5). Pebbles are dominantly of centimetre-size plates of lime mudstone and fewer but larger (10 to 40 cm) rounded clasts of grainstone and packstone. The grainy clasts are commonly intraclastic and skeletal and include trilobite, cephalopod, gastropod, crinoid and possibly some sponge fragments. Rare dolostone clasts also occur.

Plate 4. *Closeup of dolomitized carbonate conglomerate showing sandy matrix of rounded quartz grains, with 'floating' carbonate clasts.*

Plate 5. *Welded, Ordovician limestone conglomerate from the upper conglomerate unit. Large rounded grainstone clasts overlie a layer of small chips and plates of laminated lime mudstone and rare dolostone.*

BIOSTRATIGRAPHY

Trilobites were collected from several rounded, skeletal–intraclastic grainstone clasts in a number of limestone conglomerate beds in the lower conglomerate unit of the measured section and loose blocks along the measured section (Figure 3), as well as from the structurally underlying deformed section. Four clasts in the limestone conglomerates of the basal 20 metres of the section yielded a *Crepicephalus* Zone fauna comparable to a fauna collected from conglomerate clasts in the structurally underlying deformed section and from some loose blocks along the measured section. A younger fauna, suggestive of the *Dunderburgia* Zone, was collected from a conglomerate at the top of the lower conglomerate unit. One trilobite horizon having an *Elvinia* Zone fauna was located in the overlying middle slate unit. Identifiable fossils were not recovered from the upper conglomerate unit, although cross-sections of trilobites and other organisms are visible in weathered, glacially scoured outcrops.

LOWER CONGLOMERATE UNIT

The lower conglomerate unit consists of a basal subunit comprised of three beds and an upper subunit of interbedded argillite and conglomerate. Two clasts from the basal limestone conglomerate bed (12 m thick) yielded the following composite fauna:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Trilobita

?Cheilocephalus sp. undet.
Crepicephalus sp. undet.
Grandagnostus/Hypagnostus sp. undet.
Kingstonia sp. undet.

One clast from a large dislodged block of the overlying 3-m-limestone conglomerate bed provided the following taxa:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Trilobita

Crepicephalus sp. undet.
Deiracephalus sp. cf. *D. unicornis* Palmer, 1962
Holcacephalus sp. undet.
Kingstonia sp. undet.

One clast in the upper (5 m) limestone conglomerate bed yielded the trilobite *Tricrepicephalus* sp. undet.

In Alabama, *Deiracephalus unicornis* Palmer, 1962 occurs with the *Crepicephalus* Zone-correlative species *Glyptagnostus stolidotus* Öpik, 1961 in the Conasauga Formation (Palmer, 1962, page F-4, Figure 4). *Deiracephalus unicornis* Palmer, 1962 also occurs in the Cow Head Group of western Newfoundland (Palmer, 1962, page F-31; Kindle, 1982; Plate 1.3, Figure 11), in boulders collected from Zone 5 of Kindle (1982), i.e., *Cedaria* to *Crepicephalus* Zone (Table 1).

The above taxa indicate a correlation with Zone 5 of Kindle (1982), i.e., *Cedaria* to *Crepicephalus* Zone. However, based on the youngest fauna collected from these beds, the basal 20 m of the lower limestone conglomerate unit is of probable early Late Cambrian *Crepicephalus* Zone age (early Dresbachian Stage).

Structurally below the measured section, six clasts from folded and faulted conglomerates yielded well-preserved fossils of which five gave the following composite fauna:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Table 1. Correlation of rock units in the Bay of Islands area. Restricted shelf trilobite zones after Lochman-Balk and Wilson (1958), Palmer (1965), Robison (1976) and Stitt (1971, 1977, 1983). Open shelf agnostoid and polymeroid zones after Öpik (1967), Palmer (1962), Robison (1976, 1988). West Newfoundland platform trilobite zonation is provisional and subject to change: Albertan portion is based on sequence in Canada Bay (Boyce *in* Knight and Boyce, 1987); Dresbachian and Franconian parts based on faunas of the Port au Port Peninsula (Boyce, 1977, unpublished); Trempealeauan and Gasconadian intervals are based on Goose Arm sequence (Boyce *in* Knight and Boyce, 1991). Cow Head trilobite zonation after Fortey *et al.* (1982), Kindle (1982), Ludvigsen *et al.* (1989) and Young and Ludvigsen (1989). The Reluctant Head Formation and the Weasel group have been placed in the same column only to save space. Dashed lines indicate uncertainty

SYSTEM	SERIES	STAGE	TRILOBITE ZONATIONS						ROCK UNITS					
			RESTRICTED SHELF		OPEN SHELF		WEST NFLD		WEST NEWFOUNDLAND					
			POLYMEROID ZONES	POLYMEROID SUBZONES	POLYMEROID ZONES	AGNOSTOID ZONES	PLATFORM (COMPOSITE)	SLOPE (COW HEAD)	"AUTOCHTHON"		"PARAUTOCHTHON"		ALLOCHTHON	
EARLY ORDOVICIAN	CANADIAN	GASCONADIAN	SYMPHYSURINA	SYMPHYSURINA WOOSTERI			HYSTRICURUS MILLARDENSIS	SYMPHYSURINA	WATTS BIGHT FORMATION		WEASEL GROUP	UPPER CONGLOMERATE UNIT	COOKS BROOK FORMATION	
				SYMPHYSURINA BULBOSA										
				SYMPHYSURINA BREVISPICATA					BERRY HEAD FORMATION	UPPER MEMBER				LOWER ORDOVICIAN CONGLOMERATES
LATE CAMBRIAN			MISSISSQUOIA	MISSISSQUOIA TYPICALIS			MISSISSQUOIA TYPICALIS	MISSISSQUOIA TYPICALIS						
				MISSISSQUOIA DEPRESSA			PLETHOPELTIS	PHYLACTERUS SAYLESI						
	CROIXIAN	TREMPEALEAUAN	SAUKIA	EUREKIA APOPSIS										
				SAUKIELLA SEROTINA				KEITHIA SCHUCHERTI						
				SAUKIELLA JUNIA										
				RASETTIA MAGNA				KEITHIA SUBCLAVATA ONCHONOTUS RICHARDSONI						
		FRANCONIAN	SARATOGIA	DRUMASPIS TEXANA			TAENICEPHALUS	BEOTHUCKIA DUOMENTA		MIDDLE MEMBER				
				SARATOGIA LIRAE										
			TAENICEPHALUS					ROKSASPIS TURBINELLA						UPPER BEAR COVE CONGLOMERATE "A"
				PARABOLINOIDES						LOWER MEMBER				
			ELVINIA	IRVINGELLA MAJOR			CAMARASPIS CONVEXUS	ZONE 7				MIDDLE SLATE UNIT		
		DRESBACHIAN	DUNDERBERGIA				DUNDERBERGIA	ZONE 6	PETIT JARDIN FORMATION	UPPER DOLOSTONE MEMBER		LOWER CONGLOMERATE UNIT		
			PREHOUSIA				DYTREMACEPHALUS STRICTUS							
			DICANTHOPYGE											
			APHELASPIS			GLYPTAGNOSTUS RETICULATUS								
			CREPICEPHALUS			GLYPTAGNOSTUS STOLIDOTUS	CREPICEPHALUS RIVUS	ZONE 5						BRAKES COVE MEMBER
MIDDLE CAMBRIAN			CEDARIA			LEJOPYGE LAEVIGATA	WELLERASPIS NEWFOUNDLANDENSIS			MIDDLE LIMESTONE MEMBER				
	ALBERTAN		ELDORADIA		BOLASPIDELLA		ELDORADIA DUNBARI			LOWER DOLOSTONE MEMBER	RELUCTANT HEAD FORMATION			HALFWAY POINT MEMBER
			EHMANIELLA			PTYCHAGNOSTUS PUNCTUOSUS	EHMANIELLA CLOUDENSIS	ZONE 4						
						PTYCHAGNOSTUS ATAVUS		ZONE 3						
						[illegible]			MARCH POINT FORMATION					
					ORYCTOCEPHALUS	PTYCHAGNOSTUS PRAECURRENS	OLENOIDES LONGISPINUS							LOWER COOKS BROOK
			GLOSSOPLEURA			PERONOPSIS BONNERENSIS	POLYPLEURASPIS GLOSSOPLEURA WALCOTTI	ZONE 2 + ZACANTHOIDES GILBERTI	PENGUIN COVE FORMATION				IRISHTOWN FORMATION	
			ALBERTELLA				PROLIOSTRACUS							
			PLAGIURA				ALBERTOCARE (AMECEPHALUS) CLEORA							

Trilobita

agnostid gen. et sp. undet.
Cedaria sp. undet.
Coosella sp. undet.
Crepicephalus sp. undet.
Deiracephalus sp. undet.
Gen. et sp. undet.
Kingstonia sp. undet.
Tricrepicephalus murphyi Kindle, 1948
Tricrepicephalus sp. undet.

Tricrepicephalus murphyi Kindle, 1948 is also known from the Murphy Creek Formation of the Gaspé Peninsula, Québec (Kindle, 1948).

The above taxa are strongly indicative of a *Crepicephalus* Zone age. Consequently, the deformed limestone conglomerate beds are probably equivalent to the basal conglomerate unit of the measured section.

The structurally lowest limestone conglomerate in the deformed section yielded one collection containing as yet unidentified inarticulate brachiopods and agnostid and ptychopariid trilobites with possible Middle Cambrian affinity.

Clasts from a 4-m-thick conglomerate bed at the top of the lower conglomerate unit yielded a younger trilobite fauna including the following composite fauna:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Trilobita

Aphelaspis spp. undet.
Dunderbergia spp. undet.
Dytremacephalus sp. undet.
Gen. et sp. undet.
?Glyptagnostus sp. undet.
Hardyoides sp. undet.
Parahousia sp. cf. *P. subequalis* Palmer, 1965
Pseudagnostus communis (Hall and Whitfield, 1877)
ptychopariid gen. et sp. undet.
?Simulolenus sp. undet.

Parahousia subequalis Palmer, 1965 occurs in the *Dunderbergia* Zone of Nevada (Palmer, 1965, page 67). According to Shergold (1977, page 83), *Pseudagnostus communis* (Hall and Whitfield, 1877) ranges from the *Aphelaspis* Zone (late Dresbachian Stage) to the *Saukia* Zone (Trempealeauan Stage).

The above taxa indicate a correlation with Zone 6 of Kindle (1982), i.e., *Aphelaspis* to *Dunderbergia* Zone. Of the above taxa, the most common are *Aphelaspis*, *Dunderbergia* and *Dytremacephalus*. Based on the youngest contained fauna, this conglomerate is of probable Late Cambrian *Dunderbergia* Zone age (late Dresbachian Stage).

One grainstone cobble, probably from bedrock yielded the following:

Trilobita

agnostid gen. et sp. undet.
Blountia sp. undet.
Carinamala sp. undet.
Coosella sp. undet.
Coosia sp. undet.
Kingstonia sp. undet.
Komaspidella–Maryvillia sp. undet.

Another yielded the following:

Brachiopoda-Inarticulata

Gen. et sp. undet.

Trilobita

Blountia sp. undet.
?Meteoraspis sp. undet.
?Terranovella sp. undet.

The above two collections correlate with the older *Cedaria* to *Crepicephalus* Zone faunas of the lower 20-m-thick conglomeratic interval.

The trilobite fauna recovered from the lower conglomerate unit is predominantly an open-shelf type.

MIDDLE SLATE UNIT

Approximately 38 m above the base of the unit, loose weathered slabs of cleaved, dark-grey calcareous argillite and shaly limestone (derived from the immediately underlying outcrop) yielded the following:

Trilobita

Irvingella sp. cf. *I. major* Ulrich and Resser *in* Walcott, 1925–including cranidia, librigenae, thoraxes, pygidia, nearly complete specimen
pterocephaliid? sp. undet. – complete specimen

Although trilobites are extremely rare at this locality, several complete and nearly complete specimens were found. The *Irvingella* material is particularly significant because, despite the fact that it is a cosmopolitan genus, the only other species of *Irvingella* known from complete material is *Irvingella nuneatonensis* (Sharman, 1886)–see Rushton (1967, Plate 52). The presence of *Irvingella* sp. cf. *I. major* Ulrich and Resser *in* Walcott, 1925 indicates a correlation with the uppermost part of the *Elvinia* Zone and Zone 7 of Kindle (1982).

UPPER CONGLOMERATE UNIT

Conglomerates of probably Cambrian age in the upper unit are dolomitic and unfossiliferous. Overlying limestone conglomerates yielded few fossils although at 194 m above the base of the section, two clasts contained crinoid remains, and slightly higher clasts contained an unidentified porifera. The highest collection was obtained 221 m above the base of the section. Here, large boulders in dolomitic limestone conglomerate yielded cross-sections of a large, straight cephalopod and crinoid debris.

The porifera and the large cephalopod suggest that the upper limestone conglomerate unit is of Ordovician rather than Cambrian age. This is confirmed by a conodont fauna identified from a sample obtained by Dr. P.A. Cawood from the upper part of the Weasel group and identified by F. O'Brien (personal communication, 1991). It included the following conodont species:

Ansella jemtlandica (Löfgren, 1978)
Periodon aculeatus (Hadding, 1913)
Scolopodus oldstockensis Stouge, 1984

This assemblage is a Middle Ordovician late Whiterock type, which is also present in the upper part of the autochthonous Table Head Group of western Newfoundland (F. O'Brien, personal communication, 1991; see also Stouge, 1984, page 19, Figure 18).

COMPARISON WITH THE RELUCTANT HEAD FORMATION

The Reluctant Head Formation is an enigmatic unit comprising thin interbedded ribbon limestone and dolostone as well as limestone breccia/conglomerate. It may be as thick as 250 m and is distributed extensively in the eastern part of the Pasadena (NTS 12H/4) and Lomond (NTS 12H/5) map sheets, and extends down to the Humber River gorge (Gillespie, 1983; Williams *et al.*, 1982; Knight and Boyce, 1991; Knight, unpublished data; Waldron and Milne, 1991). In 1991, four new fossiliferous horizons were discovered within the unit. In addition, five of the sixteen 1990 localities (Knight and Boyce, 1991) were re-investigated. The following species have been identified:

Trilobita

Asaphiscus sp. cf. *A. wheeleri* Meek, 1873
Hemirhodon convexifrons Rasetti, 1948
Kormagnostus seclusus (Walcott, 1884)

The above fauna is dominated by pygidia of *Hemirhodon convexifrons* Rasetti, 1948, ranging in size from a couple of millimetres to at least 5 cm.

Asaphiscus wheeleri Meek, 1873 occurs in the *Bathyuriscus fimbriatus* Subzone of the *Bolaspidella* Zone in Utah (Robison, 1964, page 512, Text-figure 2).

Hemirhodon convexifrons Rasetti, 1948 was originally described from latest Middle Cambrian boulders in the Lévis Formation conglomerates of Québec (Rasetti, 1948, page 318, Table 2). A possibly related species, *Hemirhodon amplipyge* Robison, 1964, occurs in the *Bolaspidella contracta* Subzone of the *Bolaspidella* Zone (Robison, 1964, page 512, Text-figure 2) in Utah. Other species of *Hemirhodon* include the following:

Hemirhodon emarginatus Rasetti, 1946
Hemirhodon schucherti Raymond, 1937
Hemirhodon simplex Rasetti, 1946
Hemirhodon sp. of Kindle (1982; Plate 1.2, figure 18)
Hemirhodon spp. of Palmer (1968, page B45; Plate 6, figures 5-7)
Hemirhodon sp. 1 of Robison (1988, page 95; Figure 16, 1-7)
Hemirhodon vermontensis (Howell, 1937)

Of the above species, *Hemirhodon* sp. of Kindle (1982) and *Hemirhodon* spp. of Palmer (1968) occur in *Bolaspidella* Zone-correlative rocks. The remainder occur in (early Dresbachian) *Cedaria* Zone-correlative strata.

Kormagnostus seclusus (Walcott, 1884) ranges from the lower *Lejopyge laevigata* Zone (*Bolaspidella* Zone-correlative) to the *Glyptagnostus stolidotus* Zone (*Crepicephalus* Zone-correlative) (Robison, 1988, page 45).

The above taxa suggest that the top of the Reluctant Head Formation was deposited entirely within the late Middle Cambrian *Bolaspidella Zone*. Consequently, it is probably a lateral equivalent of the lower dolostone member of the Petit Jardin Formation exposed in Goose Arm. The top of the Reluctant Head Formation is thus older than the base of the Weasel group.

INTERPRETATION OF THE WEASEL GROUP

The lithofacies in the Weasel group consist of resedimented and suspension-deposited carbonates and siliciclastic mudstones. Phyllites and ribbon limestones associated with some slump beds and rare, thin lenticular limestone conglomerate, form the lowest beds in the section. The very fine grain-size and planar lamination and thin bedding suggest quiet, dominantly suspension-deposition below wave base with only minor gravity-induced instability and resedimentation. The carbonate mud was probably placed in suspension by storms stirring the bottom in the shallower parts of the ramp and shelf (see Markello and Read, 1981). This suggests that these rocks were laid down on the distal ramp slope as interpreted for similar lithofacies of the Reluctant Head Formation at Old Mans Pond to the south (Knight and Boyce, 1991).

The succeeding massive conglomerates of the lower and upper conglomerate units were probably deposited by debris flows generated at the top of the slope. This is suggested by the dominance of platy fine-grained limestones derived from the slope itself and by the presence of lesser amounts of intraclastic and skeletal grainstone clasts. The latter are similar to lithofacies in the uppermost beds of the Reluctant Head Formation that are interpreted as deposits of higher energy, shallow-water settings including a carbonate sand shoal (Knight and Boyce, 1991). The association of the debris flows with siliciclastic muds and rare calcareous beds suggests that they bypassed the lower slope and came to rest at a toe of slope–basinal setting. This, and the types of the clasts, suggest a distally steepened ramp (Tucker and Wright, 1990); this interpretation is further strengthened by the absence of clasts of true shallow-water shelf origin or clasts derived from a rim barrier, as is seen in coeval deposits of the Cow Head Group and Cooks Brook formation (James, 1981; James and Stevens, 1986; Botsford, 1988). The fossils in the lowest conglomerates indicate that they were deposited during the early and late Dresbachian *Crepicephalus* and *Dunderburgia* zones.

Predominance of siliciclastic muds with very scarce fine-grained carbonate nodules and laminae suggest that the middle slate unit was deposited in a deepwater basinal setting. Thin-bedded couplets of siltstone–shale and carbonate–shale suggest that dilute turbidites and storms may have transported fine terrigenous sediment to the basin. The calcareous laminae probably record storms that stirred up carbonate shelf mud. The trilobites at the top of the unit date the deposition of the slates as *Elvinia* Zone. A few resedimented oolitic grainstone and limestone conglomerate beds suggest renewed progradation of the shelf post-*Elvinia* Zone.

The quartz sand-rich, dolomitized conglomerates at the base of the upper conglomerate section are also probably Late Cambrian in age. Rounded quartz sand occurs in dolostones of the upper dolostone member of the Petit Jardin Formation throughout western Newfoundland (Chow, 1986; Knight, 1991; Knight and Boyce, 1991; Knight and Cawood, 1991) and is also abundant in the middle of the Tuckers Cove member of the Shallow Bay formation, Cow Head Group (James and Stevens, 1986) and the Cooks Brook formation (Botsford, 1988). James and Stevens (1986) indicate that the quartz sand reached the slope in the late Franconian to Trempealeauan. The well-rounded grains are believed to be eolian in origin and reached the deepwater by largely bypassing the shelf during sea level low-stand (Coniglio, 1985; Chow, 1986; James and Stevens, 1986).

Limestone conglomerates of the upper unit of the Weasel group revert to amalgamated, massive debris flows with insignificant fines but interbedded with dark siliciclastic shales. Dominantly lime mudstone clasts and lesser grainstone clasts rich in crinoid debris suggest the flows were generated at the upper slope and deposited at the toe of slope. Conodonts indicate that this steepened margin persisted into the Middle Ordovician.

The Reluctant Head Formation and Weasel group are preserved in two separate thrust slices but contain lithofacies of similar aspect, although different ages. This suggests that the parts of the Cambrian shelf margin from which they were structurally transported was a prograding ramp through the late Middle Cambrian (*Bolaspidella* Zone) and the early Late Cambrian (Dresbachian Stage). However, whereas the Reluctant Head Formation grades upward into peritidal shelf dolostones through a sequence of crossbedded carbonate sands (Knight and Boyce, 1991), the Weasel group is completely deepwater in origin. The predominance of conglomerates in the Weasel group suggests that the ramp at least evolved into a distally steepened type by the early Late Cambrian and continued as such throughout the Early and Middle Ordovician. The domination, through the Late Cambrian and Ordovician, of thick limestone conglomerates of very similar character to those in the Dresbachian indicates that the margin did not change significantly during the Lower and Middle Ordovician. If this is true, the Ordovician part of the Weasel group was also deposited upon a distally steepened ramp. In addition, since the Weasel group is all deepwater, the Weasel thrust slice must have originated farther basinward than the Old Mans Pond slice (Knight and Boyce, 1991) and associated thrust slices (e.g., 'Pye's ridge', see Knight, *this volume*) that carried Reluctant Head Formation and overlying shelf carbonates to their present position in the Humber tectonostratigraphic zone.

The deposition of basinal siliciclastic muds above the lower conglomerate unit suggests sea level rise drowned the margin during the early Franconian *Elvinia* Zone. The return of thick debris flows, rich in quartz sand, points to renewed shelf progradation by the end of the Franconian. The pervasive dolomitization and quartz-sand content of the conglomerates at this time seems to fit chronostratigraphically with coeval rocks of the periplatformal Cow Head Group and Cooks Brook formation. Quartz sand is abundant in the middle of the Tuckers Cove member, Shallow Bay formation, Cow Head Group and in a calcarenite interval in the Cooks Brook formation at Northern Head (James and Stevens, 1986; Botsford, 1988; personal communication, 1991). The sand reached the slope setting during the late Franconian to early Trempealeauan.

Quartz sand is also widespread but not especially abundant in peritidal shelf carbonates of the upper part of the Petit Jardin Formation, which is also pervasively dolomitized. The very fine-grained dolostones retain original sedimentary structures indicative of essentially synsedimentary dolomitization (Haywick, 1985). Since coeval dolostones also occur in the Old Mans Pond and associated thrust slices of the Pasadena and adjoining map areas, this suggests that dolomitization extended out to the shelf edge during this period. This begs the question, was the sandy carbonate conglomerate of the Weasel group dolomitized at this same time? If it was, progradation of the margin must be linked to relative sea level fall.

The conclusion that the Weasel group and Reluctant Head Formation were laid down on a Cambrian to Ordovician ramp shelf appears to differ markedly from the interpretation of the Newfoundland shelf margin deduced from the geological character of the Cow Head Group as outlined by James and Stevens (1986) and James *et al.* (1989). They showed essentially that the shelf edge adjacent to the site of the Cow Head Group was marked by carbonate sand shoals and metazoan-microbial buildups and was for the most part a rimmed to, at times, a bypass margin.

In addition, the conclusions drawn from the stratigraphy of the Weasel group and Reluctant Head Formation also appear to contrast with a recent hypothesis that a northwest-trending, cross-strike discontinuity, located at the southern end of the Long Range Precambrian inlier at Bonne Bay affected the style of sedimentation of lower Paleozoic rocks (Cawood and Botsford, 1991). Cawood and Botsford (1991) argue that there are significant differences in sedimentary facies across the proposed discontinuity. This occurs not only between the periplatform sediments of the Cooks Brook and Middle Arm Point formations of the Humber Arm Allochthon and those of the Cow Head Group, but also in the basement cover sequence, particularly the rift facies. This difference, they infer, reflects paleogeographic variations along the margin at the Newfoundland Promontory controlled by synsedimentary activity across the discontinuity. In their scheme, the Weasel group and Reluctant Head Formation should lie south of the discontinuity together with the Cooks Brook formation and Northern Arm formation.

Comparison of the Cooks Brook and the Middle Arm Point formations with the Weasel group and the Reluctant Head Formation indicates, however, that there are significant lithological and chronostratigraphic differences between the two, suggesting that no commonality existed. Studies of the Canada Bay area and the Goose Arm area show significant similarity between the Cambrian–Ordovician successions of the two areas (Knight and Boyce, 1987, 1991; Knight and Saltman, 1980; Knight, 1987). For example, as in the Goose Arm area, it is possible to trace coeval Cambrian ramp and shelf lithofacies from proximal to distal settings in an imbricate thrust stack in the Canada Bay area. This includes the presence of Reluctant Head type rocks overlain by dolomitized carbonate sands and peritidal carbonates in the most easterly thrust slices adjacent to the Hare Bay Allochthon. This implies that the ramp style of margin extended at least from the Humber Arm–Goose Arm areas northward along the eastern side of the Great Northern Peninsula. This appears to contradict the across-strike discontinuity model of Cawood and Botsford (1991). In addition, unpublished studies in 1987 by Knight and Boyce showed that the Lower to Middle Cambrian siliciclastic to carbonate succession at the south end of the Long Range Inlier records a eastward-deepening ramp similar to the eastward-deepening trend present in coeval sedimentary rocks at the northern end of the Long Range Inlier from St. Barbe to Canada Bay (Knight, unpublished data; see also James *et al.*, 1987, 1989; Knight and Cawood, 1991).

The cumulative evidence of a Cow Head margin model (rimmed platform; James and Stevens, 1986), modified rimmed to drowned-faulted margin model for the Cooks Brook–Middle Arm Point formations (Botsford, 1988; Cawood and Botsford, 1991) and the ramp model presented here for the Pasadena area indicate that the margin varied along its length.

To resolve this dilemma, it is proposed that the different style of margin (i.e., rimmed shelf versus ramp shelf) reflects deposition along different parts of a margin whose geomorphology was fundamentally structurally controlled as a result of rifting. A relatively narrow shelf having a steep-rimmed margin flanked by Cow Head type lithofacies might be expected to form along a transform margin. The westward swing of the margin across the Gulf of St. Lawrence to Gaspé, Québec, may reflect such a feature (Stockmal *et al.*, 1987; Knight *et al.*, 1991). Broad platforms having distally steepened ramps may be the margin morphology where the shelf evolves above a broad foundation of Precambrian basement such as was likely for the Newfoundland Promontory, north of the transform fault. This margin configuration requires changing the apex of the promontory as proposed by Williams (1978) and Williams and Cawood (1989) to more realistically reflect the distribution of shelf rocks and the distribution of Laurentian basement in Newfoundland inferred from Lithoprobe transects (Keen *et al.*, 1986; Marillier *et al.*, 1989). In addition, it suggests that the Taconian emplacement of the Humber Arm Allochthon was northward rather than from the east as is currently accepted (Cawood and Botsford, 1991).

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SEDIMENTOLOGICAL, GEOCHEMICAL AND SEDIMENT-PROVENANCE CONSTRAINTS ON STRATIGRAPHY AND DEPOSITIONAL SETTING OF THE STRONG ISLAND CHERT (EXPLOITS SUBZONE, NOTRE DAME BAY)

T. Dec, H.S. Swinden[1] and J.D. Floyd[2]
Newfoundland Mapping Section

ABSTRACT

The Early Ordovician volcano-sedimentary succession of the Strong Island Chert is well exposed on the southwestern shore of Strong Island in New Bay, above a conformable contact with pillow basalts of the Tea Arm Volcanics (Exploits Group). The sedimentary package of the Strong Island Chert consists of deep-water ribbon radiolarites, graptolitic black shales, felsic fallout tuffs and volcaniclastic debrites and turbidites. The sediments are interbedded with a number of units of basalt pillow lava and basalt massive flow.

The preliminary geochemical data of the pillow lavas from the upper part of the Tea Arm Volcanics display signatures of island-arc tholeiites. The overlying basalts of the Strong Island Chert have signatures of alkaline basalts and ocean-island tholeiites and are interpreted to have erupted in a back-arc basin. The observed arc-to-back-arc, evolutionary nature of the entire volcano-sedimentary succession provides the basis for a provisional chemostratigraphic correlation of the Exploits Group with the Wild Bight Group.

The volcaniclastic debrites and turbidites constitute a significant component of the Strong Island Chert. They are characterized by the abundance of felsic, andesitic and mafic detritus and together with the felsic fallout tuffs are interpreted to have been transported into the back-arc basin from an active island arc. An additional continent- and ophiolite-derived sediment input may be represented by polymictic turbidites containing a metamorphic and chromite detritus.

INTRODUCTION

The New Bay sector of Notre Dame Bay (Figure 1) contains the type area of the Exploits Group (following the usage of Dean, 1977 and Kean *et al.*, 1981), which forms the substrate on which the Strong Island Chert was deposited. The unit has been subdivided into four mappable units of formational rank (Helwig, 1967; O'Brien, 1990). In ascending order these are: the Tea Arm Volcanics, the Saunders Cove Formation, the New Bay Formation and the Lawrence Head Volcanics (O'Brien, 1991) (Figure 2). The maximum combined stratigraphic thickness of the formations of the Exploits Group occurs in the east of the New Bay area, where at least 3.5 km of strata underlies the Strong Island Chert.

The Strong Island Chert, although relatively thin (around 200 m thick), is regionally extensive and has been provisionally designated by O'Brien (1990) as a mappable unit of a formational status (Figure 2). It rests conformably in different localities on the lower and the upper Tea Arm Volcanics, the upper New Bay Formation and the top of the Lawrence Harbour Volcanics. The Strong Island Chert passes conformably upward into graptolitic (Caradocian) black shales of the Lawrence Harbour Shale (O'Brien, 1991).

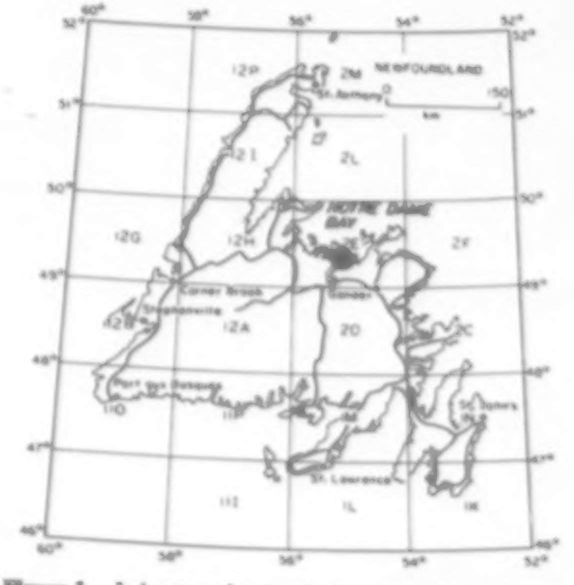

Figure 1. *Index map showing the location of the study area.*

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[1] Mineral Deposits Section, Newfoundland Department of Mines and Energy, Geological Survey Branch

[2] British Geological Survey, Murchison House, West Mains Road, Edinburgh, EH9 3LA, Scotland

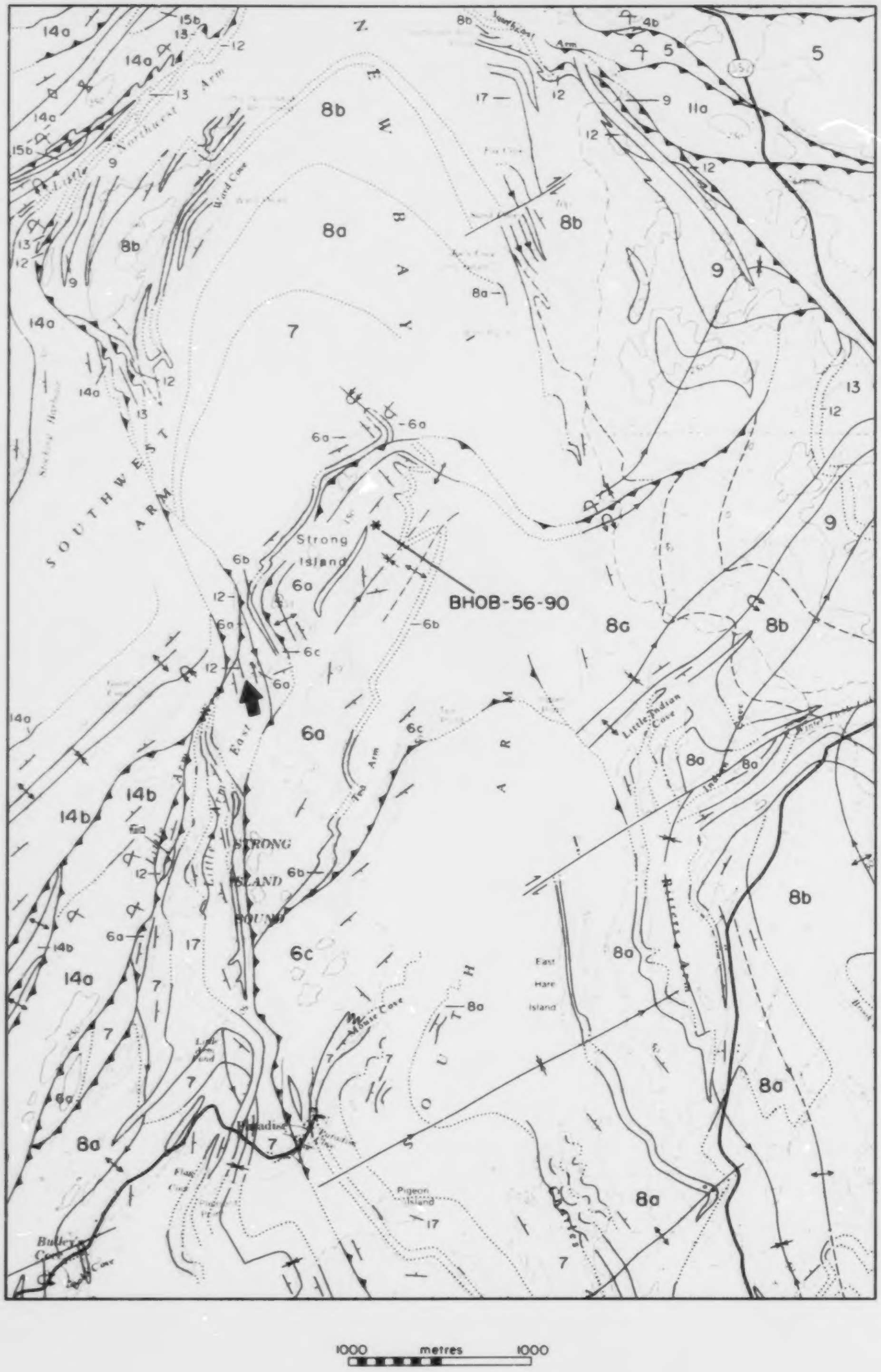

NEW BAY
SOUTHWEST ARM
SOUTH ARM
Little Northwest Arm
Strong Island
BHOB-56-90
STRONG ISLAND SOUND
East Arm
Tea Arm
Little Indian Cove
Rattling Arm
Northeast Arm
East Hare Island
Pigeon Island
Paradise
Burley's Cove
14a
14b
15b
13
12
9
8a
8b
7
6a
6b
6c
5
4b
11a
17
1000 metres 1000

Figure 2. *Geological map of the New Bay area (modified part of the open file map by O'Brien, 1990). The bold arrow points to the contact of the Strong Island Chert with the Tea Arm Volcanics and the location of the type section (see Figure 3).*

LEGEND (for Figure 2)

SILURIAN OR DEVONIAN

SOUTH ARM GABBRO

17 *Dykes, sills and sheets of gabbro, diorite and associated diabase and mafic pegmatite*

ORDOVICIAN AND SILURIAN

GOLDSON CONGLOMERATE

15b *Poorly sorted and stratified, grey, feldspathic and siliceous wacke*

ORDOVICIAN

POINT LEAMINGTON GREYWACKE

14b *Dark grey, carbonaceous shale and light grey, quartzofeldspathic wacke; olistostrome having a black shale matrix*

14a *Interbedded grey shale and light grey, quartzofeldspathic wacke*

LAWRENCE HARBOUR SHALE

13 *Unseparated units of black carbonaceous shale; black pyritiferous siltstone having black shale partings; grey chert having bioturbated black argillite laminae*

STRONG ISLAND CHERT

12 *Interbedded, radiolarian chert, mafic pillow lavas and reddish-brown feldspathic wacke; minor, green and red, laminated argillite; rare felsic tuff*

COBBS ARM LIMESTONE

11a *Tectonized equivalents within melange belts of the Boones Point Complex*

ORDOVICIAN OR EARLIER

EXPLOITS GROUP

LAWRENCE HEAD VOLCANICS

9 *Mafic pillow lavas*

NEW BAY FORMATION

8b *Feldspathic wacke and grey argillite; minor conglomeratic to pebbly wacke*

8a *Thin-bedded grey shale; interbedded grey shale and sandy wacke; conglomeratic to pebbly wacke (lower part of unit)*

SAUNDERS COVE FORMATION

7 *Red chert; interbedded red and green, siliceous argillite; feldspathic wacke; minor conglomerate*

TEA ARM VOLCANICS

6c *Mafic pillow lavas, pillow breccia and agglomerate; minor feldspathic wacke and laminated argillite*

6b *Felsic tuff and agglomerate*

6a *Mafic pillow lavas*

COTTRELLS COVE GROUP

MOORES COVE FORMATION

5 *Thick-bedded to thin-bedded, reddish-brown feldspathic wacke; grey argillite*

FORTUNE HARBOUR FORMATION

4b *Dark green, mafic pillow lavas*

KEY

Geological boundary (defined, approximate, assumed)..........

Bedding (inclined, overturned)..........

Anticline (upright, overturned)..........

Syncline (upright, overturned)..........

Fold axial trace (plunge direction indicated)..........

Antiform..........

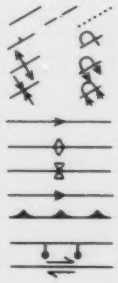
Synform..........

Sideways-closing fold..........

Thrust fault (teeth on upthrown side)..........

Normal fault (solid circle on downthrown side)..........

Strike-slip fault..........



NOTE

For the sake of simplicity, foliation symbols are purposefully omitted from all map units.

Numerical designations of rock formations are the same as in the Open File Map 90-124 by O'Brien (1990).

The Middle–Upper Ordovician Strong Island Chert and Lawrence Harbour Shale together form an overstep sequence onlapping various formations of the Exploits and Wild Bight groups (O'Brien, 1991). The Strong Island Chert is nonetheless in conformable contact with all these formations. The lack of a surface of erosional unconformity beneath the Strong Island Chert (the conformable contact with the Tea Arm Volcanics is marked with arrow in Figure 2) demonstrates that the considerable thickness changes of the constituents of the Exploits Group near the New Bay Fault Zone are an original depositional feature of the Early–Middle Ordovician sedimentation, rather than an artifact of later (Silurian) tectonic removal of parts of the succession (O'Brien, 1991).

The Strong Island Chert is composed of radiolarian cherts, siliceous shales, felsic tuffs (containing so far undated zircons and conodonts), black shales (containing early Llanvirn graptolites; see Williams *et al.*, *this volume*) and epiclastic sandstones and conglomerates (Figures 3 and 4). This remarkably heterogeneous sedimentary assemblage is interbedded with a number of pillowed and massive basalt flows. The type section of the Strong Island Chert is probably the most continuous, time-controlled and geochemically constrained record of Llanvirn–Llandeilo sedimentation and magmatism in central Newfoundland.

This report establishes sedimentological, geochemical and sediment–provenance constraints on the depositional setting of the Strong Island Chert. This information, as well as an understanding of the relationship with the underlying Exploits Group, is vital to drawing correlations between the Exploits and Wild Bight groups and deciphering the geology of the Exploits Subzone.

TYPE SECTION OF THE STRONG ISLAND CHERT

The type section of the Strong Island Chert is very well exposed on the southwestern shore of Strong Island, above a conformable contact with pillow lavas of the Tea Arm Volcanics (arrowed in Figure 2). The succession of north striking and near vertical sedimentary rocks and interbedded lavas is little deformed and a continuous section, 127 m thick, has been measured through the main portion of the Strong Island Chert (Figures 3 and 4). A pumpellyite–prehnite-facies mineral assemblage has been identified throughout the measured section. In this report, the rocks of the Strong Island Chert are subdivided into three main groups (Figure 4):

1) SILICEOUS DEPOSITS

The most characteristic rocks of the Strong Island Chert, constituting approximately 51 percent of the measured section, are brown-red and grey to green and turquoise-green radiolarian cherts together with associated siliceous brown-red, grey, green and black shales ('ribbon radiolarites', Plate 1a). Those cherts that occur directly above and close to the lavas have a distinctive turquoise colour, caused by abundant celadonite.

The grey and green cherts are commonly bioturbated. The burrows are seen in sections normal to the bedding as Fe-enriched, wispy mottles. The non-bioturbated cherts and shales show erosional bases, normal grading, parallel lamination and small-scale crosslamination that indicate deposition from low-density turbidity currents and from pelagic fallout. Early Llanvirn graptolites have been recovered from one black shale horizon (Figure 4; see Williams *et al.*, *this volume*). Parallel-laminated felsic tuffs and volcaniclastic sandy turbidites are a minor constituent of the siliceous deposits of the Strong Island Chert. Minor sedimentary slumping and bedding-parallel tectonic disruption (thinning–thickening of the section) is mostly confined to the assemblage of ribbon radiolarites.

2) VOLCANICLASTIC DEPOSITS

Volcaniclastic turbiditic sandstones and conglomerates and debris-flow conglomerates (or debrites) constitute around 19 percent of the measured section. The turbidites represent a broad spectrum of low- to high-density currents, some of which are transitional to debris flows. The debrites are matrix-supported, structureless and extremely poorly sorted. The gravel fraction of the volcaniclastic deposits is represented mainly by intrabasinal fragments of siliceous sediments that reach up to 50 cm in diameter. Some of the debris-flow beds display normal grading and have erosional bases, cutting up to 2 m deep into the underlying sediments (Figure 4).

One isolated debris-flow conglomerate consists of basalt clasts up to 30 cm in size set in a fine-grained siliceous matrix. This massive debrite is sharply underlain and overlain by the siliceous fine-grained deposits (Figure 4).

3) BASALT LAVAS

Seven massive and pillowed lava flows make up approximately 30 percent of the measured section. One massive unit consists of at least three amalgamated flows, separated locally by thin, fine-grained siliceous deposits and volcaniclastic sandstones (see the complementary section in Figure 4). The upper and lower contacts of the lavas with the sediments are sharp, with no evidence of brecciation or pyroclastic fragmentation. Locally, the underlying sediments have been squeezed and bulldozed by the overlying pillows (Plate 1b).

GEOCHEMICAL SIGNATURES OF THE LAVAS

Pillow lavas present a difficult problem in correlation for the field geologist because basalts that record distinctly different magmatic events may look very similar in the field. It has been shown in several areas of central Newfoundland that geochemical data from mafic volcanic rocks can provide important information about the nature and correlation of volcanic units, both on a local and regional scale (e.g., Swinden 1987; Swinden *et al.*, 1989; Dunning *et al.*, 1991; Swinden and Jenner, *this volume*). Of particular interest to this study are geochemical data for the Wild Bight Group

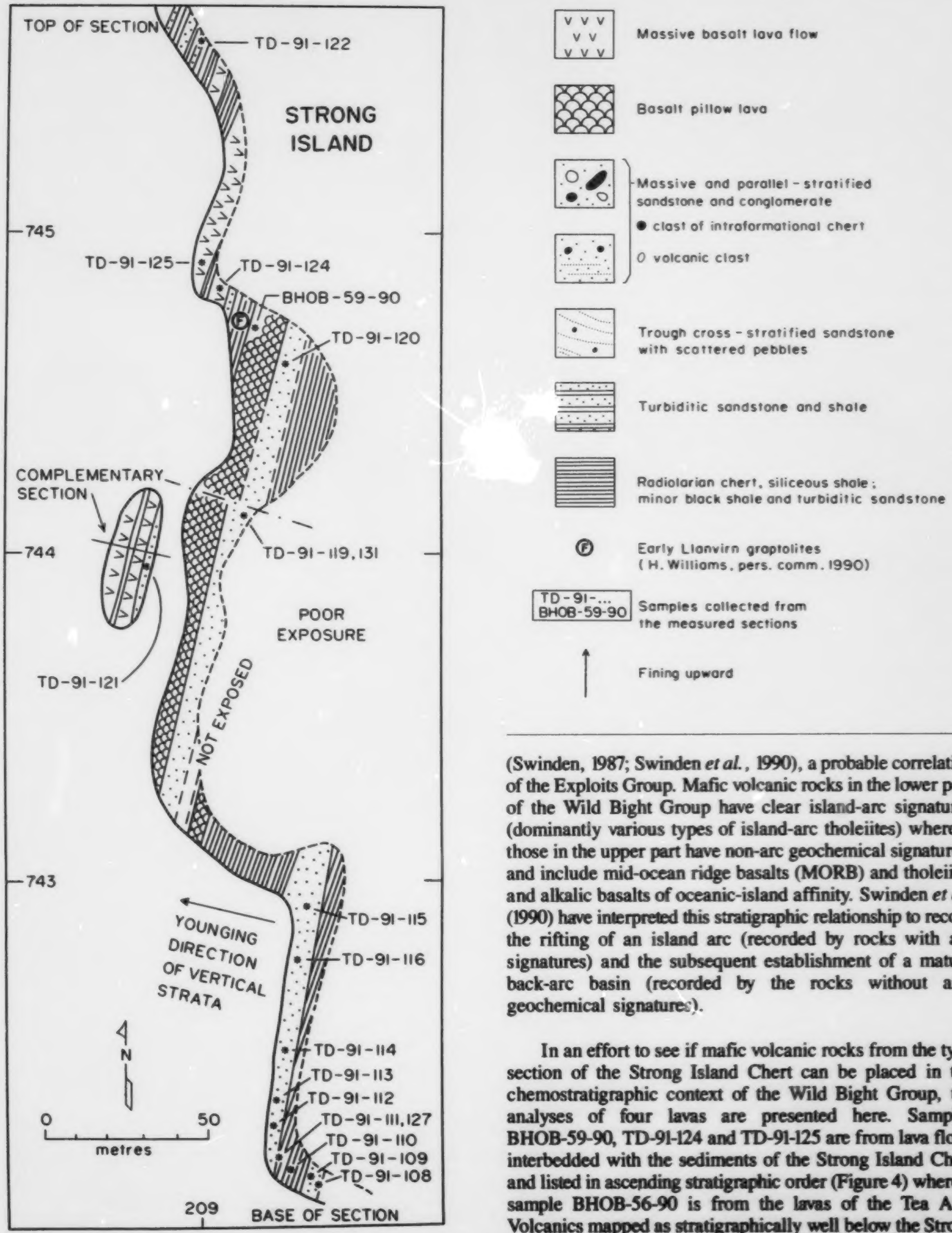

TOP OF SECTION
TD-91-122
STRONG ISLAND
745
TD-91-125
TD-91-124
BHOB-59-90
TD-91-120
COMPLEMENTARY SECTION
744
TD-91-119, 131
POOR EXPOSURE
NOT EXPOSED
TD-91-121
743
YOUNGING DIRECTION OF VERTICAL STRATA
TD-91-115
TD-91-116
N
TD-91-114
TD-91-113
TD-91-112
TD-91-111, 127
TD-91-110
TD-91-109
TD-91-108
0
50
metres
209
BASE OF SECTION
Massive basalt lava flow
Basalt pillow lava
Massive and parallel - stratified sandstone and conglomerate
● clast of intraformational chert
O volcanic clast
Trough cross - stratified sandstone with scattered pebbles
Turbiditic sandstone and shale
Radiolarian chert, siliceous shale; minor black shale and turbiditic sandstone
Early Llanvirn graptolites (H. Williams, pers. comm. 1990)
TD-91-... BHOB-59-90
Samples collected from the measured sections
Fining upward

Figure 3. *Detailed map of the measured portion of the type section exposed on the southwestern shore of Strong Island.*

(Swinden, 1987; Swinden *et al.*, 1990), a probable correlative of the Exploits Group. Mafic volcanic rocks in the lower part of the Wild Bight Group have clear island-arc signatures (dominantly various types of island-arc tholeiites) whereas those in the upper part have non-arc geochemical signatures, and include mid-ocean ridge basalts (MORB) and tholeiitic and alkalic basalts of oceanic-island affinity. Swinden *et al.* (1990) have interpreted this stratigraphic relationship to record the rifting of an island arc (recorded by rocks with arc signatures) and the subsequent establishment of a mature back-arc basin (recorded by the rocks without arc-geochemical signatures).

In an effort to see if mafic volcanic rocks from the type section of the Strong Island Chert can be placed in the chemostratigraphic context of the Wild Bight Group, the analyses of four lavas are presented here. Samples BHOB-59-90, TD-91-124 and TD-91-125 are from lava flows interbedded with the sediments of the Strong Island Chert and listed in ascending stratigraphic order (Figure 4) whereas sample BHOB-56-90 is from the lavas of the Tea Arm Volcanics mapped as stratigraphically well below the Strong Island Chert (Figure 2). We have complete major and routine trace-element data for two samples, and trace-element data only for the other two (Table 1). These data were acquired

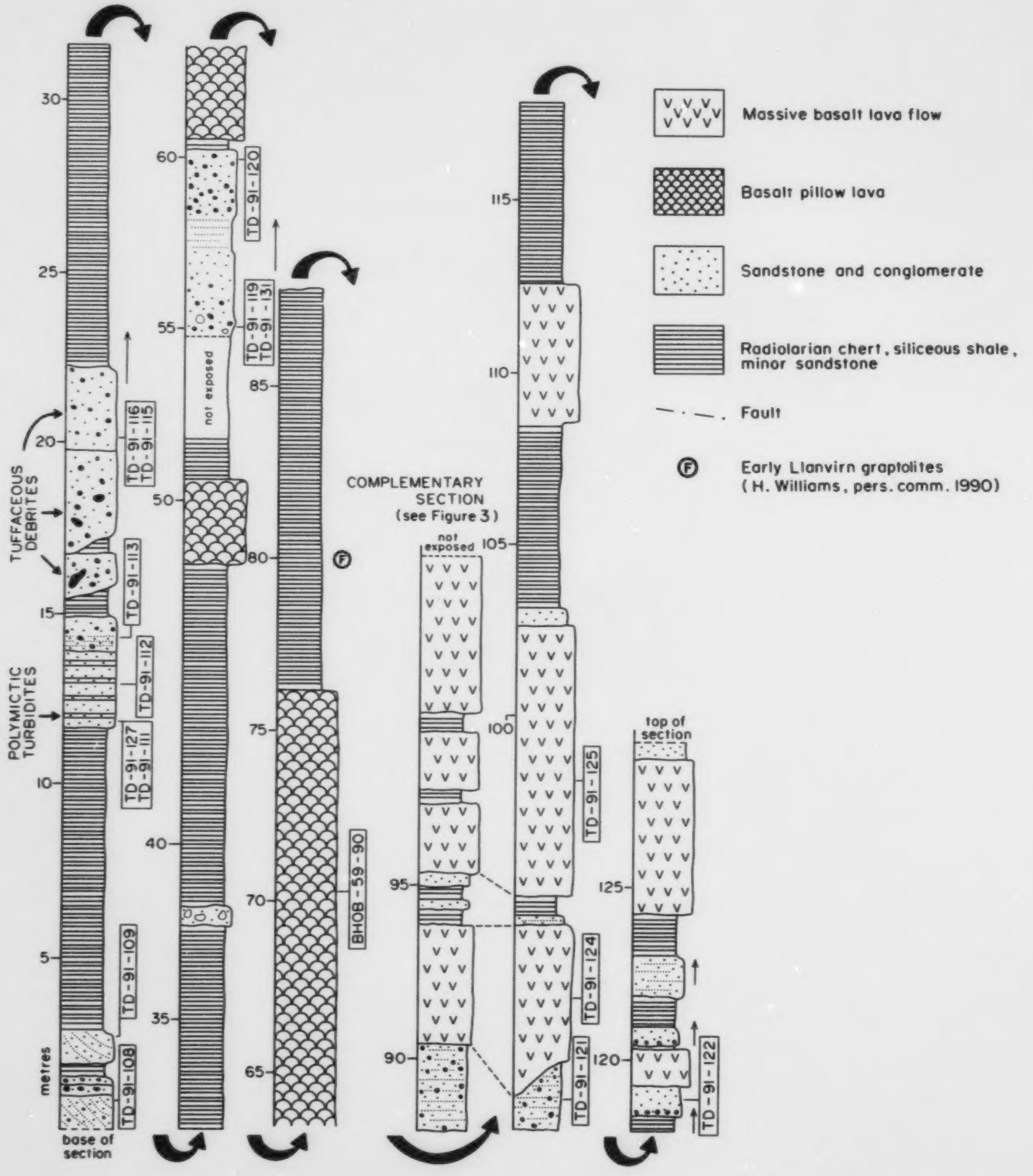

Massive basalt lava flow
Basalt pillow lava
Sandstone and conglomerate
Radiolarian chert, siliceous shale, minor sandstone
Fault
Ⓕ Early Llanvirn graptolites (H. Williams, pers. comm. 1990)
TUFFACEOUS DEBRITES
POLYMICTIC TURBIDITES
metres
base of section
top of section
not exposed
COMPLEMENTARY SECTION (see Figure 3)
TD-91-108
TD-91-109
TD-91-127
TD-91-111
TD-91-112
TD-91-113
TD-91-116
TD-91-115
TD-91-119
TD-91-131
TD-91-120
BHOB-59-90
TD-91-125
TD-91-124
TD-91-121
TD-91-122
5
10
15
20
25
30
35
40
50
55
60
65
70
75
80
85
90
95
100
105
110
115
120
125

Figure 4. *Graphic log of the measured continuous section of the Strong Island Chert.*

by inductively coupled plasma–emission spectrometry (ICP-ES) in the Geological Survey Branch Laboratory using HF, $HClO_4$ and HCl digestion. In addition, we have a suite of high-field-strength element (HFSE) and rare earth-element (REE) data for two samples, analyzed by inductively coupled plasma–mass spectrometry (ICP-MS) at the Department of Earth Sciences, Memorial University. Techniques for accuracy and precision for the ICP-MS analyses are reported by Jenner *et al.* (1990).

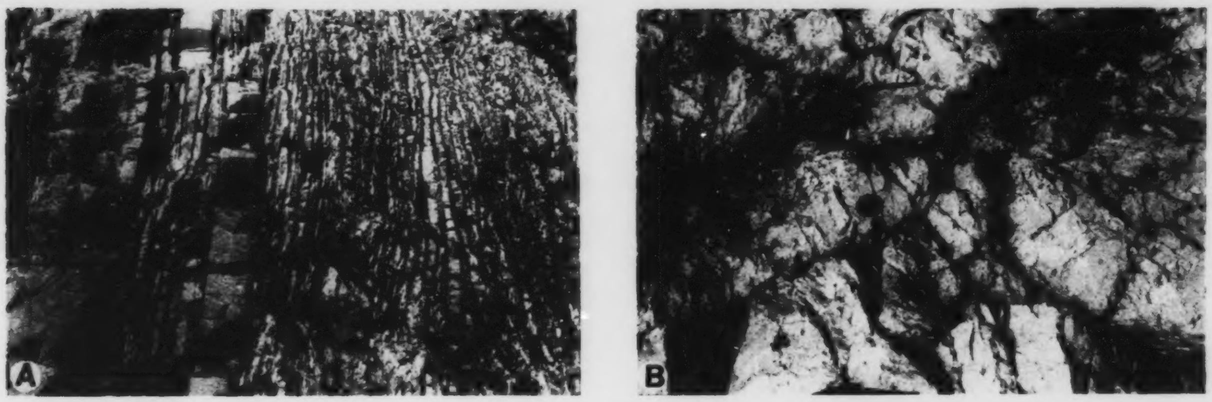


Plate 1. *(a) Ribbon radiolarites overlain by polymictic sandstone turbidites; the turbidites (samples TD-91-111, 127) contain cryptic fragments of phyllite, schist, quartzite detritus and accessory zircon, tourmaline, garnet, magnetite, chromite and epidote. Lens cap is 5 cm wide. (b) Pillow lavas squeezing underlying chert.*

The two samples for which there are both HFSE and REE data show a clear contrast in their geochemical signatures (Figure 5). The sample from the Tea Arm Volcanics (BHOB-56-90) displays light-rare-earth-element (LREE) depletion and a prominent negative Nb anomaly on a primitive mantle normalized basis. It is very similar to the Wild Bight Group Type A-II lavas (Swinden *et al.*, 1990), which are interpreted as island-arc tholeiites occurring in about the middle of the stratigraphy and below any lavas of non-arc affinity. Sample BHOB-59-90 from the Strong Island Chert, in contrast, is a strongly LREE-enriched basalt that lacks the arc-geochemical signature. It is similar to the Wild Bight Group Type N-I lavas, interpreted as mildly alkalic basalts that occur only in the upper part of the stratigraphic section. The geochemical dissimilarity of these samples is further emphasized by the disparity in ratios of more-incompatible to less incompatible elements, illustrated here on the Ti-V diagram (Figure 6). The relatively low Ti/V ratio of the Tea Arm Volcanics sample BHOB-56-90 is typical of arc basalts whereas the very high ratio of sample BHOB-59-90 is typical of alkalic basalts. In the two samples for which only trace-element data are available, the TiO_2 contents are a little high for most Wild Bight Group arc rocks but are not in themselves diagnostic. However, the ratios of more-incompatible to less-incompatible elements are higher than is typical of arc basalts (see Figure 6). These rocks appear to be geochemically similar to Type N-II basalts in the Wild Bight Group (Swinden *et al.*, 1990), which are interpreted to be tholeiites of oceanic-island affinity, although more complete data are required to confirm this interpretation. Such rocks characteristically occur together with alkalic basalts in the Wild Bight Group.

Although it is speculative to extrapolate too much from a limited analytical base, the geochemical data for the four samples from the southwestern shore of Strong Island do suggest some preliminary conclusions. The geochemical relationships are consistent with those observed in the Wild Bight Group. Probable oceanic-island tholeiites and alkalic basalts occur high in the section whereas island-arc thoeliites occur lower down. This provides some support to previous suggestions that the Wild Bight and Exploits groups are stratigraphically related. The association of alkalic basalts with high-TiO_2 tholeiites in the Strong Island Chert is similar to the association observed in the inferred back-arc section of the Wild Bight Group (Swinden *et al.*, 1990) and this, coupled with the stratigraphic correlations, suggests that the Strong Island Chert may have formed in the same back-arc basin as the upper part of the Wild Bight Group.

COMPOSITIONAL FEATURES OF THE VOLCANICLASTIC DEPOSITS

The volcaniclastic deposits of the Strong Island Chert carry additional information about the depositional context of the volcano-sedimentary succession. Reconnaissance fieldwork and preliminary petrographic investigations lead to a subdivision of the volcaniclastic deposits into four provenance categories. Except for one basalt-derived conglomerate, the volcaniclastic rocks have no direct affinity with the intercalated mafic lava flows.

The above mentioned *basalt-derived conglomerate* occurs in the lower portion of the measured section as a single, 0.5-m-thick debris-flow bed (Figure 4). It consists of angular basalt cobbles and pebbles surrounded by a fine-grained matrix.

Tuffaceous sandstones occur as turbidites and as the matrix material in both the debrites (TD-91-113, -115, -116; Figure 4) and the trough cross-stratified conglomerates (TD-91-108, -109; Figure 4). An abundance of highly vesicular, vitric, wispy fragments of predominantly pyroclastic origin (5.7 to 44.3 percent) suggests that there was little reworking of the tuffaceous sands. The vitric material is typically altered to chlorite, sphene, epidote, albite, prehnite and calcite, and represents mostly mafic and intermediate compositions. The tuffaceous sandstones contain abundant felsic detritus (felsitic texture) as well as fragments of andesite (identified through microlitic texture) and basalt (lathwork texture) (e.g.,

Table 1. Geochemical analyses of selected samples of basalts from the Strong Island Chert and Tea Arm Volcanics

	BHOB-56-90	BHOB-59-90	TD-91-124	TD-91-125
SiO_2	58.65	55.95	-	-
Al_2O_3	12.47	12.89	-	-
Fe_2O_3	1.98	1.47	-	-
FeO	9.45	10.36	-	-
MgO	4.26	4.67	-	-
CaO	6.36	5.14	-	-
Na_2O	4.27	5.18	-	-
K_2O	0.93	0.13	-	-
TiO_2	1.28	3.06	-	-
MnO	0.20	0.37	-	-
P_2O_5	0.16	0.79	-	-
LOI	7.85	4.79	-	-
TOTAL	99.08	99.22	-	-
Mg#	40.33	41.61	-	-
Cr	8	15	-	-
Ni	7	4	-	-
Sc	28	24	-	-
V	411	157	257	342
Ta	0.09	2.74	-	-
Nb	1.5	48	19	13
Hf	2.12	8.34	-	-
Zr	66	341	249	286
Y	24	44	38	47
Th	0.98	6	-	-
La	4.80	41.10	-	-
Ce	13.38	94.29	-	-
Pr	1.95	12.06	-	-
Nd	10.07	51.83	-	-
Sm	3.27	11.59	-	-
Eu	0.99	3.90	-	-
Gd	3.84	11.87	-	-
Tb	0.38	1.65	-	-
Dy	4.81	9.60	-	-
Ho	1.03	1.78	-	-
Er	3.11	4.74	-	-
Tm	0.44	0.64	-	-
Yb	3.20	4.02	-	-
Lu	0.46	0.58	-	-

– no data; major elements in wt %; trace elements in ppm

Dickinson, 1970; Ingersoll and Cavazza, 1991). The proportion of embayed volcanogenic quartz grains ranges from 8.7 to 44.0 percent (Table 2). Many volcanic rock fragments show some degree of Fe-oxide alteration dating from before the release as sediment into the basin.

The tuffaceous sandstones contain a significant proportion (4.0 to 9.0 percent, Table 2) of slightly corroded, sand-size fragments of brachiopods, trilobites, sponges, algae, crinoids, bryozoa and possibly corals.

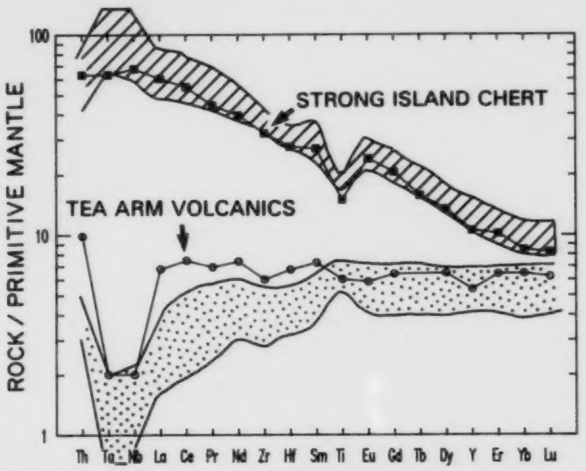


Figure 5. *Extended rare-earth plots for samples BHOB-56-90 (open circles) and BHOB-59-90 (closed squares). Note the prominent negative Nb and Ta and positive Th anomalies for BHOB-56-90, the characteristic island-arc geochemical signature (Swinden* et al., *1990). Stippled field is Type A-II, LREE-depleted island-arc tholeiites from the Wild Bight Group; hatched field is Type N-I alkalic basalts from the Wild Bight Group. Primitive mantle-normalization values are given by Swinden* et al. *(1990) and are approximately twice chondrite.*

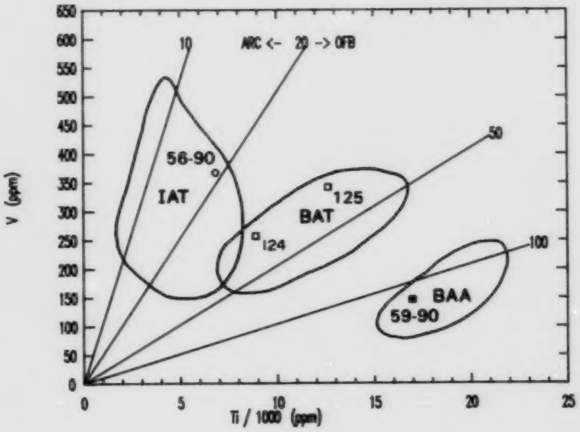


Figure 6. *The Ti-V diagram after Shervais (1982) illustrating the contrast in ratios of more-incompatible to less-incompatible elements in the four samples from Strong Island. Fields are from the Wild Bight Group (Swinden, 1987): IAT–island-arc tholeiites; BAT–tholeiites of back-arc origin, including MORB-like rocks and rocks of oceanic-island affinity; BAA–back-arc alkalic basalts.*

Reworked volcanogenic sandstones have been found in two turbidite beds (TD-91-119, -120 and -121). The sandstones are coarse grained, well sorted, grain-supported and contain only trace amounts of vitric detritus. A secondary cement

Table 2. Mineralogy of detrital constituents in the volcaniclastic deposits of the Strong Island Chert

Sample	TD-91-113	TD-91-115	TD-91-116	TD-91-120	TD-91-119	TD-91-121	TD-91-111
PETROFACIES	Tuffaceous Sandstones			Reworked Sandstones			Polymictic Sandstones
Monocrystalline quartz	21	13	8.7	44.0	47.3	36.1	29.7
Plagioclase	9.3	21.9	10.3	33.7	42.3	50.5	44.3
ROCK FRAGMENTS							
Felsic	28.0	15.0	17.3	11.7	4.0	8.0	15.3
Andesite	14.0	7.0	7.7	2.7	2.3	4.3	1.3
Basalt	6.0	12.6	4.7	2.0	1.0	0.3	0.3
Bioclasts	9.0	4.0	5.7	–	–	–	0.3
Quartz arenite	–	–	–	–	–	–	2.3
Other rock fragments	0.0	1.3[1]	–	–	–	0.7	0.7[1]
Vitric	11.0	6.6	44.3	5.7	1.3	–	–
Tuffaceous shale	0.3	18.3	1.3	0.3	0.6	–	–
ACCESSORY MINERAL GRAINS							
Zircon	tr	–	–	tr	tr	tr	tr
Tourmaline	–	–	–	–	–	–	tr
Garnet	–	–	–	–	–	–	tr
Hornblende	–	–	–	–	–	tr	–
Muscovite	–	–	–	–	–	tr	tr
Biotite	tr	tr	tr	tr	tr	tr	5.0
Epidote	tr	–	–	–	–	tr	tr
Chromite	–	–	–	–	–	tr	tr
Magnetite	tr	tr	tr	tr	tr	tr	tr
Ilmenite	tr	tr	–	–	–	–	–
Apatite	–	–	–	–	–	tr	tr
Matrix/cement	48.0	62.9	38.0	28.9	16.7	13.1	48.4

[1] phyllite–schist, tr–trace, – no data

of pumpellyite–prehnite mineral assemblage fills the pore spaces. Abundant volcanogenic quartz (36.1 to 47.3 percent) and albitized plagioclase (33.7 to 50.5 percent) are accompanied by mafic to felsic volcanic-rock fragments (Table 2).

Polymictic sandstones are restricted to thin turbidites in the lower part of the measured section (TD-91-111, -127; Plate 1a). The sandstones are poorly sorted but contain no vitric material. Apart from the ubiquitous volcanogenic detritus the polymict sandstones contain fragments of phyllite, phyllite-schist, quarzite and quartz arenite, as well as an unusually abundant detrital biotite. Although some of the biotite is present as phenocrysts in the felsic volcanic-rock fragments, there are common grains of kinked biotite, which might have been derived from metamorphic schists. Detrital zircon, tourmaline (occurring as a detrital constituent of the quartz arenite clasts), garnet, epidote, magnetite, chromite and apatite occur as accessory components (Table 2).

DEPOSITIONAL SETTING OF THE STRONG ISLAND CHERT

Despite the fact that the Deep Sea Drilling Project has failed to find a modern analog for the ancient biogenic cherts (e.g., Hein and Karl, 1983), there is an agreement among researchers that the ribbon radiolarites are deep-marine deposits that accumulated in small, arc-related, marginal basins (Jenkyns and Winterer, 1982; Hein and Karl, 1983; Jones and Murchey, 1986).

A substantial proportion of the Strong Island Chert probably formed in an oxic environment, assuming that the metabolism of the radiolarian ooze-burrowing organisms was based on oxygen. The abundance of corroded bioclastic material found in the tuffaceous debrites (e.g., TD-91-115, -116; Table 2) suggests that the basin floor was, at least temporarily, either above or not far below the carbonate compensation depth.

Examination of the association of radiolarian cherts with ophiolites appears most instructive when addressing the problem of depositional environment (Jenkyns and Winterer, 1982; Jones and Murchey, 1986) because many ophiolites represent relicts of arc-related basins. The geochemical data and the correlation with the Wild Bight Group strongly suggest that the volcano-sedimentary succession of the Strong Island Chert formed in a back-arc basin.

The paleotectonic model for the upper part of the Wild Bight Group takes into account a possibility of concurrently operating island-arc and back-arc volcanism (Swinden *et. al.*, 1990, p. 235). The tuffaceous debrites of the Strong Island Chert, which are dominated by sand-size felsic, andesitic and basaltic detritus having a significant pyroclastic component (samples TD-91-113, 115, -116; Figure 4, Table 2), must have been derived from an active volcanic arc (e.g., Klein, 1985; Nishimura *et al.*, 1991) rather then from the intercalated submarine basalts. The postulated island-arc volcanism would also explain the felsic tuff bands and the reworked, quartz-enriched, epiclastic sediments. The scoured bases of the tuffaceous debrites imply that the Strong Island Chert was relatively proximal to the site of the debris-flow mobilization (e.g., Pierson, 1980; Prior *et al.*, 1984).

Only one basalt debrite might have been derived from lava flows, similar to those present within the section. Geochemical signatures of the basalt clasts will be examined to confirm or reject such a provenance.

The emplacement of volcaniclastic debris flows and turbidites was most likely linked to a copious supply of pyroclastic ejecta from the island arc. The accumulation of such an arc-derived volcaniclastic aprons would have been essentially independent of relief and tectonic uplift (Klein, 1985; but see also Carey and Sigurdson, 1984).

Unlike the felsic tuffs in the Wild Bight Group (Swinden *et al.*, 1990), those in the Strong Island Chert contain abundant euhedral zircon. The inferred island-arc eruptions are, therefore, distinctly different from those recorded by the felsic tuffs of the Wild Bight Group.

Polymictic sandstone turbidites in the lower part of the section (TD-91-111, -127, Figure 4 and Plate 1A) contain fragments of phyllite-schist, quartzite and quartz arenite, accompanied by exceptionally abundant detrital biotite and trace garnet (Table 2). This exotic material may have been derived from rocks of continental affinity. The accessory chromite indicates a minor sediment input from an eroded ophiolite. Detritus of continental (Gander Zone) and ophiolitic (Gander River Complex) provenance has also been found in the late Arenig–early Llanvirn sandstone of the Weir's Pond Formation (basal unit of the Davidsville Group), which rests unconfomably on the ultramafic rocks of the Gander River Complex, and in the time-equivalent sandstone of the Indian Bay Big Pond Formation (O'Neill, 1991; Wonderley and Neuman, 1984; Boyce *et al.*, 1988; see also Dec and Colman-Sadd, 1990; Colman-Sadd *et al.*, *in press*).

CONCLUSIONS

Preliminary sedimentological and geochemical data suggest that the type section of the Strong Island Chert contains a record of Early Ordovician deep-water sedimentation in a back-arc basin that accompanied a coeval island-arc volcanism. The geochemical signatures of the basalt flows and pillow lavas from the Strong Island Chert and from the underlying Tea Arm Volcanics support a provisional correlation of the Exploits Group with the arc-to-back-arc succession of the Wild Bight Group.

The deposition of the felsic tuffs and the volcaniclastic debrites and turbidites was controlled by a sediment supply from an active island arc. Local polymictic sandstone turbidites are characterized by a metamorphic detritus of a possible continental provenance and detrital chromite derived from ophiolite.

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OPHIOLITES, SEDIMENTARY ROCKS, POSTTECTONIC INTRUSIONS AND MINERALIZATION IN THE EASTERN POND (NTS 2D/11W) MAP AREA, CENTRAL NEWFOUNDLAND

W.L. Dickson, P.Geo.
Newfoundland Mapping Section

ABSTRACT

The Eastern Pond (NTS 2D/11W) map area contains components of the Exploits and Mount Cormack subzones that are juxtaposed along the northern margin of the ophiolitic Great Bend Complex. A newly discovered zone of amphibolite schist and mylonitic gabbro represents a basal dynamothermal aureole to the ophiolite. This zone marks the junction of the suite with the Lower Ordovician or older Spruce Brook Formation and the Silurian Botwood Group. Possible Silurian sedimentary rocks are imbricated with the ultramafic rocks.

In the eastern part of the area, the polydeformed turbiditic slate and sandstone of the Middle to Late Ordovician Davidsville Group are possibly overlain by turbiditic, shallow-marine and intertidal sandstone and siltstone of the Silurian Botwood Group. Upper Ordovician graptolite-bearing slate of the Davidsville Group forms a fault-bound block within the Botwood Group. Upper Ordovician graptolite-bearing chert and slate are possibly equivalent to the Lawrence Harbour Formation. Shallow-marine to terrestrial Botwood Group sedimentary rocks underlie the northwestern part of the area and these are in fault contact with the Spruce Brook Formation.

The Mount Peyton Intrusive Suite is composite and consists of postkinematic gabbro that is cut by tonalite, granodiorite and granite. The suite has intruded the Spruce Brook Formation and the Botwood Group. Isotopic dating of the pegmatitic gabbro indicates a Silurian age. Two other layered gabbroic units are possibly related to the suite rather than being part of the ophiolite.

Postkinematic, very thick-bedded, mainly coarse-grained, sedimentary rocks, probably of fluvial origin, unconformably overlie fossiliferous Upper Ordovician chert, the Botwood Group, and a small part of the ophiolite suite.

The southern half of the Great Bend Complex is extensively altered to magnesite and gold mineralization is concentrated in this zone. Massive arsenopyrite mineralization occurs in brecciated serpentinite. The Chiouk Brook showing contains anomalous gold values. Antimony mineralization occurs in graphitic slate of the Davidsville Group and also the postkinematic sandstone and conglomerate at the Xingchang and Hunan deposits, respectively.

INTRODUCTION

LOCATION AND ACCESS

The Eastern Pond (west half) map area (formerly known as the West Gander Rivers map area) is located in central Newfoundland, approximately 40 km southeast of Grand Falls. During 1991, the area was mapped using 1:12 500 colour aerial photographs taken during 1976, 1978 and 1987. Much of the map area is accessible by a network of woods roads, constructed since 1981 (see Figure 1; also Dickson, 1991 for detailed map), from Route 360, which runs along the western edge of the map area. The easternmost portion of the map area can be reached via woods roads from the Trans Canada Highway at Glenwood in map area NTS 2D/15. Road access and the main roads, as of 1987, are shown on the 1:250 000 map (NTS 2D, Edition 3). (It should be noted that the bridge over the Northwest Gander River, 12 km southwest of Gander Lake, has been omitted from this map). The Northwest Gander River may be canoed in the spring but much outcrop is covered by water at this time. By July, in most years, water levels are very low and the river can then be crossed on foot. Helicopter support is available at Gander, Glenwood, Milltown and St. Alban's.

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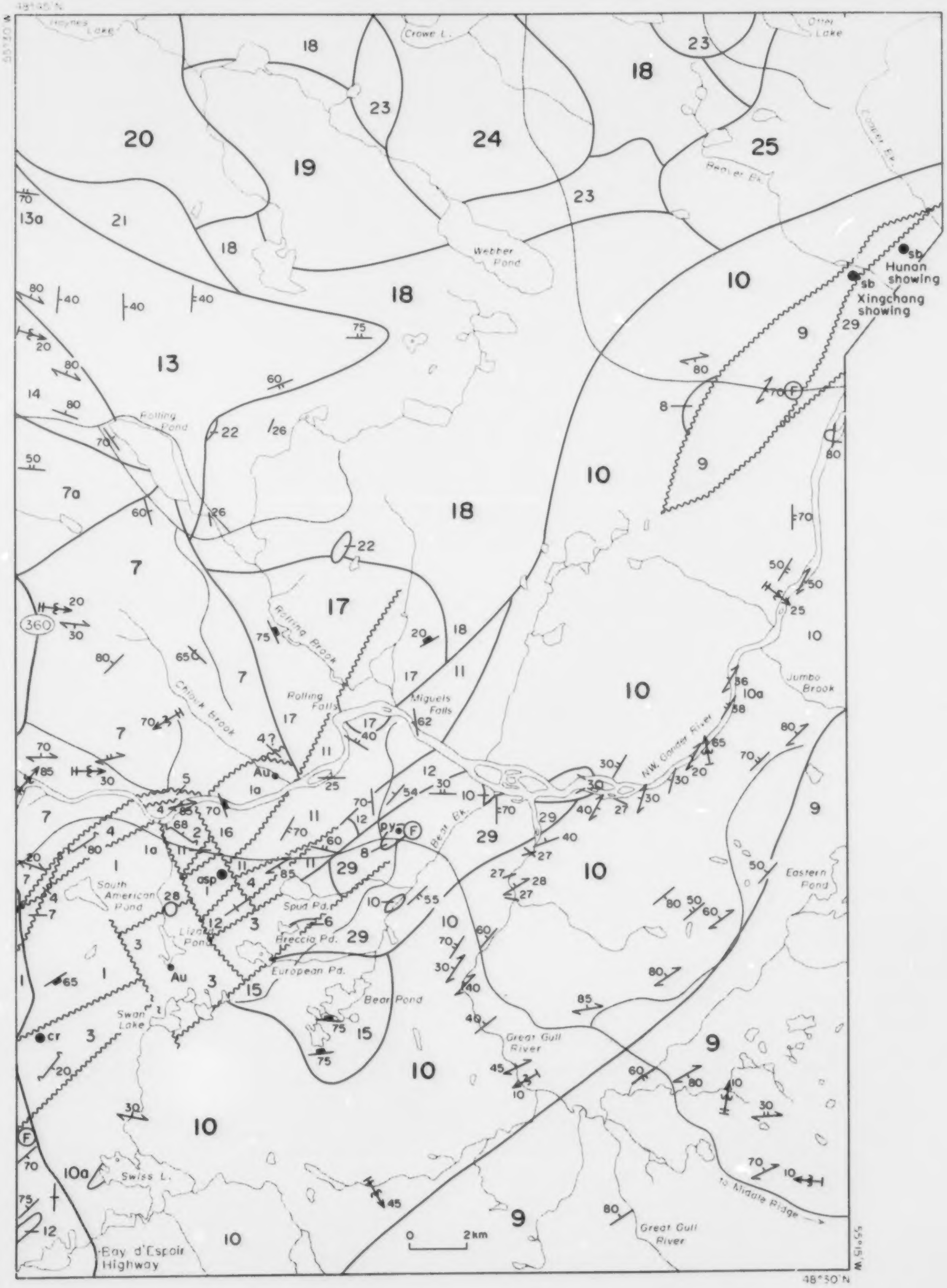

48°45'N
55°30'W
Haynes Lake
Crowe L.
Otter Lake
Beaver Bk.
Webber Pond
Hunan showing
Xingchang showing
sb
Rolling Pond
Rolling Brook
Rolling Falls
Miguels Falls
Chiuck Brook
360
Au
asp
py
cr
South American Pond
Lizard Pond
Spud Pd.
Breccia Pd.
European Pd.
Bear Pond
Swan Lake
Swiss L.
Bay d'Espoir Highway
Bear Bk.
NW. Gander River
Jumbo Brook
Eastern Pond
Great Gull River
to Middle Ridge
0
2km
48°30'N
55°15'W

Figure 1. *Geological map of the Eastern Pond (NTS 2D/11W) map area.*

LEGEND

DEVONIAN or YOUNGER

30 *Very thick-bedded, coarse breccia with imbricated slate clasts (derived from Unit 10) and well-rounded polymict pebbles and cobbles*

29 *Very thick-bedded, feldspathic sandstone, siltstone, pebble–cobble conglomerate, red, brown and black shale; contains conspicuous chert clasts derived from Unit 8*

28 *Peridotite-cobble to pebble conglomerate and minor peridotite-clast sandstone*

SILURIAN to DEVONIAN

27 *Fine- to medium-grained diabase dykes*

MT. PEYTON INTRUSIVE SUITE (UNITS 17 TO 26)

26 *Fine- to medium-grained diabase dykes*

25 *Medium- to fine-grained, locally miarolitic, leucocratic biotite granite*

24 *Fine-grained, equigranular, biotite granite*

23 *Medium-grained, equigranular, hornblende–biotite granodiorite*

22 *Fine-grained, equigranular, biotite–hornblende tonalite*

21 *Medium-grained, equigranular, leucocratic, muscovite–biotite granite locally displaying a strong eutaxitic foliation; commonly contains xenoliths of migmatized sediment probably from Unit 13*

20 *Medium-grained, hornblende–biotite gabbro*

19 *Medium-grained, hornblende gabbro cut by abundant granitic dykes and veins*

18 *Medium- to fine-grained, hornblende gabbro; locally contains very coarse pegmatitic hornblende gabbro*

17 *Medium- to coarse-grained, layered to uniform, hornblende gabbro*

16 *Medium- to coarse-grained, layered to uniform, hornblende gabbro*

15 *Coarse-grained, hornblende gabbro and diabase; gabbro locally displays a weak mineral alignment*

SILURIAN

NORTH STEADY POND FORMATION AND BOTWOOD GROUP (Units 10 to 14)

14 *Thin- to thick-bedded, well-cleaved, red and green siltstone, sandstone, conglomerate and breccia*

13 *Buff-weathering, grey, mudcrack-bearing, boudinaged, sandstone hornfels;* **13a,** *migmatized equivalents*

12 *Red and green, weakly cleaved, siltstone and sandstone locally containing mud cracks and rain pits*

11 *Grey, mudcrack-bearing, sandstone hornfels*

10 *Grey to green, well-cleaved, mudcrack-bearing sandstone and siltstone, slate, coarse-grained sandstone and pebbly sandstone, rare calcareous, fossiliferous sandstone; minor sandstone hornfels; thin-bedded, turbiditic sandstone;* **10a,** *pebble conglomerate*

MIDDLE TO LATE ORDOVICIAN

9 Davidsville Group *Polydeformed, greywacke sandstone, black graptolitic slate, granule conglomerate*

8 *Black, pyritic, ribbon chert and minor graptolitic black slate*

EARLY ORDOVICIAN ?

7 Spruce Brook Formation *Polydeformed, greywacke sandstone, siltstone, quartzose sandstone and slate;* **7a,** *thick-bedded, quartz-rich sandstone and siltstone*

CAMBRIAN to EARLY ORDOVICIAN ?

GREAT BEND COMPLEX (Units 1 to 6)

6 *Silicified, red ultramafic breccia*

5 *Chloritized basalt and schistose mafic tuff*

4 *Amphibolite and chloritized amphibolite schist, probably of gabbroic origin, and mylonitic serpentinite*

3 *Mainly schistose magnesite containing minor chromite layers and pods*

2 *Massive to schistose, layered dunite pyroxene and harzburgite*

1 *Massive to sheared, layered to uniform, serpentinized peridotite; includes dunite and minor harzburgite;* **1a,** *Sheared serpentinite–shale or slate melange*

SYMBOLS

Geological contact (not classified, approximate, assumed)	——— — — — ....
Bedding (tops known, unknown, overturned)	—⊥— —⊥⊥— —⊕—
Primary igneous layering or mineral alignment	—●—
Cleavage (first, second; inclined, vertical)	↔ ↔ ↔ ↔
Shear or mylonitic fabric (inclined vertical)	^—v ^—+—v
Fault (not classified)	〰〰
Fold axis (first, second generation)	⊢→ ‖→
Dyke	▬▬▬
Mineral occurrences (gold, arsenopyrite, chromite, pyrite, magnetite, stibnite)	Au,as,cr,py,mt,sb
Fossil locality	F

Since 1982, the areas that were opened up for logging have been clear-cut for pulp wood. Logging continues in the Webber Pond and Jumbo Brook areas. Most of the cut-over areas and intervening areas of poor-quality trees, west of a line from Haynes Lake to Great Gull River, were ravaged by a forest fire in 1986. Consequently, much more bedrock is now exposed than was available to the pre-1983 geologists, e.g., Colman-Sadd (1982). Extensive portions of the map area are still covered by poor-quality spruce trees, bogs and deadfall.

REGIONAL GEOLOGICAL SETTING

Most of the area lies within the Exploits Subzone of the Dunnage Zone, which is dominated by Ordovician sedimentary and volcanic units and Ordovician ophiolite complexes. The eastern part of the Mount Cormack Subzone, which is included in the Gander Zone (Williams *et al.*, 1988), lies along the western part of the map area. The subzone has been interpreted as lying within a thrust-bound window within the Exploits Subzone (Colman-Sadd and Swinden, 1984) with the fault contacts commonly marked by ultramafic rocks. The Mount Cormack Subzone is dominated by Lower Ordovician and older sedimentary rocks.

The ophiolitic rocks within the map area (Units 1 to 6; Figures 1 and 2) form the eastern portion of an extensive, fragmented ophiolite complex that extends well to the west (see Colman-Sadd and Swinden, 1982, 1984; Swinden, 1988; Dec and Colman-Sadd, 1990). The ultramafic rocks and adjacent gabbro in the map area have been informally termed the Great Bend (± Ophiolite or Ultramafic) Complex (or body). The abrupt change in course of the Northwest Gander River at the mouth of Miguels Brook was informally termed Big Bend by Grady (1953) and then renamed Great Bend by Harrison (1953).

PREVIOUS WORK

The work of Grady (1952, 1953) was the first significant geological work in the area. He outlined the ultramafic and gabbroic rocks and described their relationships to the adjacent sedimentary rocks. The contacts were known to be faulted but were interpreted to have been originally intrusive. The sedimentary rocks were correlated with the Ordovician sedimentary rocks in the Hamilton Sound area of Notre Dame Bay and the igneous rocks were interpreted to be have been intruded during the Ordovician Taconic Orogeny (Grady, 1953, page 40). The extensive deposits of high-quality magnesite within the ultramafic rocks in the Great Bend area were emphasized in his report. Grady *(op. cit.)* also informally named the minor ponds in the area i.e., Lizard Pond, Swiss Lake, South American Pond, Spud Pond, European Pond, and Bear Pond. These names are used on most geological maps of the area (Figure 2).

The ultramafic rocks have received considerable attention from exploration companies interested initially in their magnesite and asbestos potential and more recently in their precious-metal potential. The few, more academic studies have concentrated on the regional significance of the ultramafic and adjacent rocks. The sedimentary and plutonic rocks in the area have received little detailed examination.

Kean (1974) examined the Great Bend Complex and concluded that the ultramafic and gabbroic rocks had intruded the adjacent sedimentary rocks as mantle diapirs. Malpas and Strong (1975) examined chrome−spinel (chromite) crystals from allochthonous ophiolitic complexes of western Newfoundland and from the ultramafic rocks in eastern Newfoundland (cf. Kean, 1974). They suggested that the difference in the Cr:Al ratios of the chrome−spinel from eastern and western Newfoundland supported the diapir hypothesis.

Colman-Sadd and Swinden (1982, 1984) interpreted the Great Bend Complex to be a continuation of the transported ophiolite complex exposed to the southwest at Coy Pond (see e.g., Colman-Sadd, 1981).

Regional mapping at a scale of 1:250 000 was carried out by Anderson and Williams (1970). Colman-Sadd (1982) did reconnaissance 1:50 000-scale mapping in the Rolling Pond−Swan Lake areas. Zwicker and Strong (1986) mapped the Great Bend ophiolitic and adjacent rocks in the southwestern part of the map area. The present study has produced the first 1:50 000-scale map of NTS 2D/11 (west half) map area (Dickson, 1991) and complements the work of Blackwood (1981) who mapped the NTS 2D/11 (east half) map area. Adjacent map areas were mapped by Blackwood (1983, 2D/6) and Colman-Sadd and Russell (1988, 2D/12). The Mount Peyton (NTS 2D/14) map area will be mapped during 1992.

DESCRIPTION OF THE UNITS

GREAT BEND COMPLEX (UNITS 1 TO 6)

Introduction

The term Great Bend Complex is used here for the ultramafic rocks (Units 1, 2, 3 and 6; Plates 1 and 2), along with the newly discovered, intensely deformed, amphibolitic gabbroic and amphibolite schists (Unit 4; Plate 3) at the northern margin of the Complex, and the possibly related, mafic volcanic rocks of Unit 5 (Figure 2). It appears, however, that the gabbroic rocks at Bear Pond and also 2.5 km northwest of Spud Pond, respectively, (Units 15 and 16), may not form a continuous area of gabbro or be genetically related to the ultramafic rocks as is indicated by Grady (1953), Kean (1974), Colman-Sadd and Swinden (1982, 1984) and Colman-Sadd (1982). The ultramafic complex is now generally considered to be part of a dismembered ophiolite suite (Colman-Sadd, 1982).

Age

The age of the Great Bend Complex has not been determined. However, the probably correlative Pipestone Pond Complex (Swinden, 1988) has been dated by Dunning and Krogh (1985). Zircon from plagiogranite in the Complex gave a U–Pb age of $493^{+2.5}_{-1.9}$ Ma.

Stratigraphic evidence for the minimum age of the Great Bend Complex comes from the adjacent Mount Cormack Subzone where the Spruce Brook Formation is unconformably overlain by fossiliferous limestone conglomerate that contains chromite grains and serpentine clasts (Dec and Colman-Sadd, 1990). The fossils include brachiopod, trilobite and other bioclastic material (see also Colman-Sadd and Swinden, 1984; Boyce, 1987) that indicate an age between the late Llanvirn and early Llandeilo (approximately 468 Ma). The Spruce Brook Formation is therefore dated as Arenig or older. The presence of the ultramafic-derived detritus in this unit indicates that the ophiolite was exposed and subject to erosion by this date (see Dec and Colman-Sadd (1990) for a full discussion).

A sample of amphibolite from the northern margin of the ophiolite is currently being dated by the Ar–Ar method and this may give an estimate of the age of emplacement of the Complex.

Contact Relationships

All observed contacts of the Great Bend Complex and adjacent units are either brittle faults or shear zones (Figure 2). The southeastern contact of the Complex (Unit 3 against Units 10 and 15) is a brittle fault, which juxtaposes sheared magnesite against a variety of rocks. Siltstone of the North Steady Pond Formation is hydrothermally brecciated and silicified, 2 km northeast of Swiss Lake. Two km west of Bear Lake, medium-grained gabbro is brecciated but not quartz-veined. At Breccia Pond, highly fractured and hematized and silicified breccia (Unit 6) is interpreted to be of ultramafic origin (Figure 2). Highly sheared, chloritic serpentinite is exposed along the road, 1.5 km northeast of Breccia Pond. The shear surfaces are generally steeply dipping.

The northwestern contact has been significantly redefined compared to all previously published work (cf. Colman-Sadd, 1982; Zwicker and Strong, 1986). This is the result of the discovery of widespread amphibolite schist, mylonitic serpentinite and coarse-grained, amphibole-bearing strongly foliated metagabbro, chlorite schist (Plate 3), and a melange composed of shale and altered ultramafic rocks (subunit 1a and Unit 4). These rocks occur within a northeast-trending band, up to 1 km wide, which extends along the previously mapped northwestern boundary of the Complex from the Bay d'Espoir Highway to the Northwest Gander River and eastward to Chiouk Brook (Figure 2; see Burton, 1987 and Mercer, 1988b).

Small outcrops of definite sedimentary rocks occur as apparently fault-bound slivers within the amphibolitic unit. At the western end of the zone, the sedimentary rocks of the Spruce Brook Formation contain well-defined mineral bands possibly reflecting mylonitization. Immediately north of the amphibolitic rocks, the grey, cleaved sandstone and siltstone of the Spruce Brook Formation are complexly deformed and contain conspicuous chlorite, whereas farther north, chlorite is not apparent in hand specimen.

Approximately 1 km northwest of Spud Pond, several outcrops of amphibolite, chlorite schist and banded mylonitic serpentinite (Unit 4) form another northeast-trending block of amphibolite schist and mylonitic serpentinite. This block is apparently displaced to the southeast of the main area of Unit 4 along a northwest-trending fault. These rocks are in fault contact with weakly cleaved red siltstone and grey to buff sandstone of the Botwood Group (Units 10 and 11; Figure 2).

These two areas of intensely deformed and highly metamorphosed rocks are interpreted to be a possible *basal dynamothermal aureole to the Great Bend Complex*)see Williams and Smyth, 1973).

A series of northwest-trending faults, west of Spud Pond, juxtapose various components along the northeastern margin of the Complex with undeformed, layered to uniform gabbro (Unit 16) and the Botwood Group sedimentary rocks (Units 10, 11 and 12). Within the Complex, slivers of sandstone and siltstone, which are part of the Botwood Group, are interpreted to be thrust-bound slices.

A recently released company report (Mercer, 1988b) listed drill-core data from the mineralized area within the Complex, south of Lizard Pond. Here, extensive outcrops of magnesite and silicified magnesite are commonly surrounded by glacial float that contains abundant siltstone clasts. Drilling indicated that the ultramafic rocks were small thrust-bound bodies enveloped in highly sheared or fractured siltstone and shale along with minor sandstone and conglomerate. The ultramafic bodies have apparent thicknesses of about 5 to 75 m and the thicker ultramafic intersections commonly contain shears and schistose bands. Commonly, the ultramafic and sedimentary rocks form imbricated slices. The greatest reported intersection of ultramafic rocks is just over 50 m and this drillhole was terminated in talc–magnesite schist (see Mercer, 1988b for details).

Unit Descriptions

The Great Bend Complex has been divided into 6 main units. Unit 1 is mainly composed of fine-grained massive, layered or uniform, dark-brown weathering, black peridotite (Plate 1). The dominant variety of peridotite is uniformly-textured dunite. On weathered surfaces, chromite is conspicuous as 1- to 2-mm-diameter crystals and up to 1 percent in abundance.

Toward the margins of this unit, the rocks are intensely sheared with a well-developed schistosity (subunit 1a). The sheared contact of peridotite and Unit 3 (magnesite) is well exposed in a trench 1 km east of South American Pond. The mineralogical changes are gradational over 10 m. This is

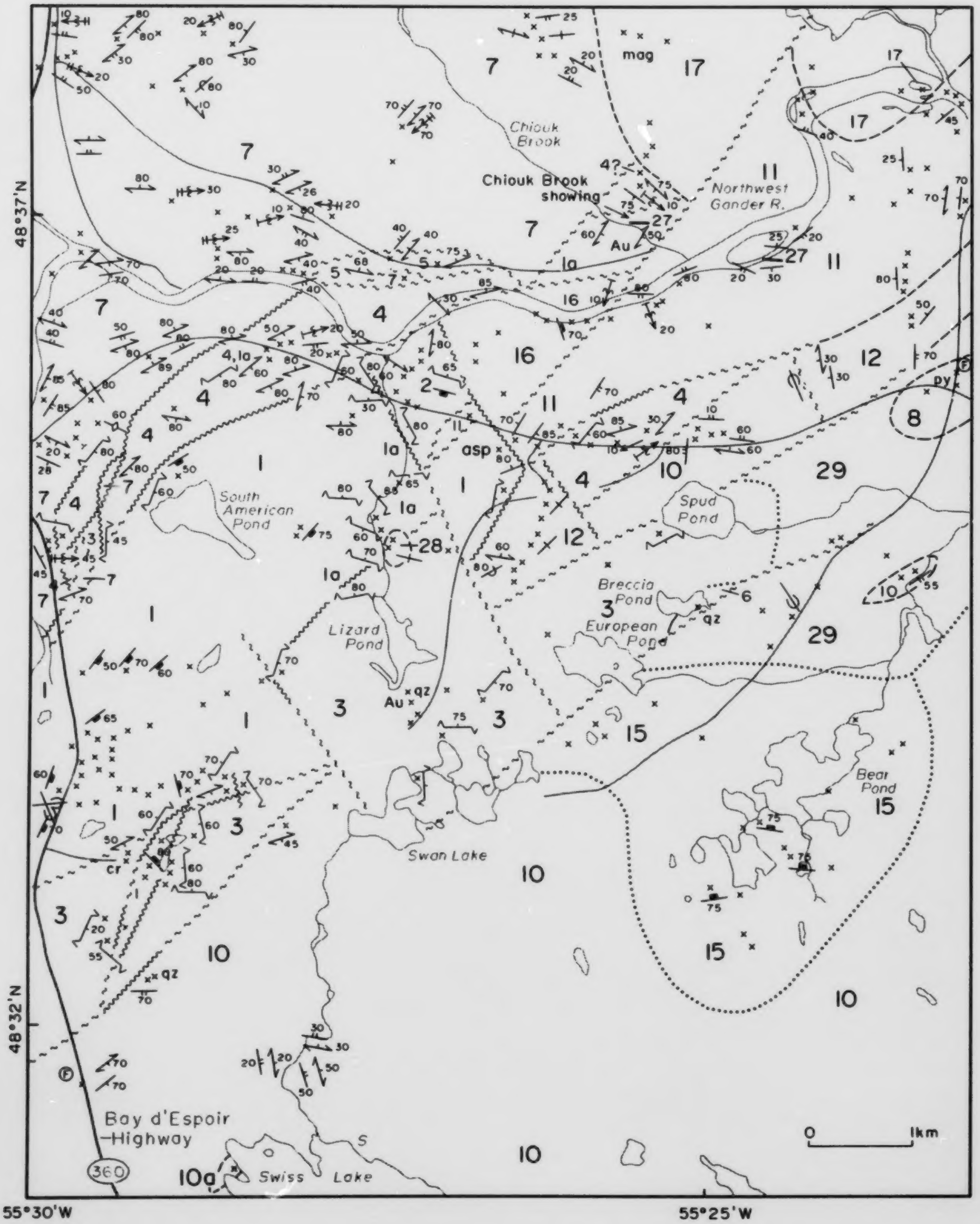

48°37'N
48°32'N
55°30'W
55°25'W
Chiouk Brook
Chiouk Brook showing
Northwest Gander R.
mag
Au
asp
py
qz
cr
South American Pond
Lizard Pond
Spud Pond
Breccia Pond
European Pond
Bear Pond
Swan Lake
Swiss Lake
Bay d'Espoir Highway
360
0
1km
7
17
11
27
5
4
16
12
8
29
1
1a
2
3
4,1a
28
6
10
10a
15

Figure 2. *Geological map of the Great Bend Complex in the Eastern Pond map-area.*

LEGEND

DEVONIAN or YOUNGER

30 *Very thick-bedded, coarse breccia with imbricated slate clasts (derived from Unit 10) and well-rounded polymict pebbles and cobbles*

29 *Very thick-bedded, feldspathic sandstone, siltstone, pebble–cobble conglomerate, red, brown and black shale; contains conspicuous chert clasts derived from Unit 8*

28 *Peridotite-cobble to pebble conglomerate and minor peridotite-clast sandstone*

SILURIAN to DEVONIAN

27 *Fine- to medium-grained diabase dykes*

MT. PEYTON INTRUSIVE SUITE (UNITS 17 TO 26)

26 *Fine- to medium-grained diabase dykes*

25 *Medium- to fine-grained, locally miarolitic, leucocratic biotite granite*

24 *Fine-grained, equigranular, biotite granite*

23 *Medium-grained, equigranular, hornblende–biotite granodiorite*

22 *Fine-grained, equigranular, biotite–hornblende tonalite*

21 *Medium-grained, equigranular, leucocratic, muscovite–biotite granite locally displaying a strong eutaxitic foliation; commonly contains xenoliths of migmatized sediment probably from Unit 13*

20 *Medium-grained, hornblende–biotite gabbro*

19 *Medium-grained, hornblende gabbro cut by abundant granitic dykes and veins*

18 *Medium- to fine-grained, hornblende gabbro; locally contains very coarse pegmatitic hornblende gabbro*

17 *Medium- to coarse-grained, layered to uniform, hornblende gabbro*

16 *Medium- to coarse-grained, layered to uniform, hornblende gabbro*

15 *Coarse-grained, hornblende gabbro and diabase; gabbro locally displays a weak mineral alignment*

SILURIAN

NORTH STEADY POND FORMATION AND BOTWOOD GROUP (Units 10 to 14)

14 *Thin- to thick-bedded, well-cleaved, red and green siltstone, sandstone, conglomerate and breccia*

13 *Buff-weathering, grey, mudcrack-bearing, boudinaged, sandstone hornfels;* **13a**, *migmatized equivalents*

12 *Red and green, weakly cleaved, siltstone and sandstone locally containing mud cracks and rain pits*

11 *Grey, mudcrack-bearing, sandstone hornfels*

10 *Grey to green, well-cleaved, mudcrack-bearing sandstone and siltstone, slate, coarse-grained sandstone and pebbly sandstone, rare calcareous, fossiliferous sandstone; minor sandstone hornfels; thin-bedded, turbiditic sandstone;* **10a**, *pebble conglomerate*

MIDDLE TO LATE ORDOVICIAN

9 Davidsville Group *Polydeformed, greywacke sandstone, black graptolitic slate, granule conglomerate*

8 *Black, pyritic, ribbon chert and minor graptolitic black slate*

EARLY ORDOVICIAN ?

7 Spruce Brook Formation *Polydeformed, greywacke sandstone, siltstone, quartzose sandstone and slate;* **7a**, *thick-bedded, quartz-rich sandstone and siltstone*

CAMBRIAN to EARLY ORDOVICIAN ?

GREAT BEND COMPLEX (Units 1 to 6)

6 *Silicified, red ultramafic breccia*

5 *Chloritized basalt and schistose mafic tuff*

4 *Amphibolite and chloritized amphibolite schist, probably of gabbroic origin, and mylonitic serpentinite*

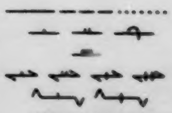
3 *Mainly schistose magnesite containing minor chromite layers and pods*

2 *Massive to schistose, layered dunite pyroxene and harzburgite*

1 *Massive to sheared, layered to uniform, serpentinized peridotite; includes dunite and minor harzburgite;* **1a**, *Sheared serpentinite–shale or slate melange*

SYMBOLS

Geological contact (not classified, approximate, assumed)
Bedding (tops known, unknown, overturned)
Primary igneous layering or mineral alignment
Cleavage (first, second; inclined, vertical)
Shear or mylonitic fabric (inclined vertical)
Fault (defined, approximate)
Fold axis (first, second generation)
Dyke
Mineral occurrences (gold, arsenopyrite, chromite, pyrite, magnetite, stibnite) Au,as,cr,py,mt,sb
Fossil locality F



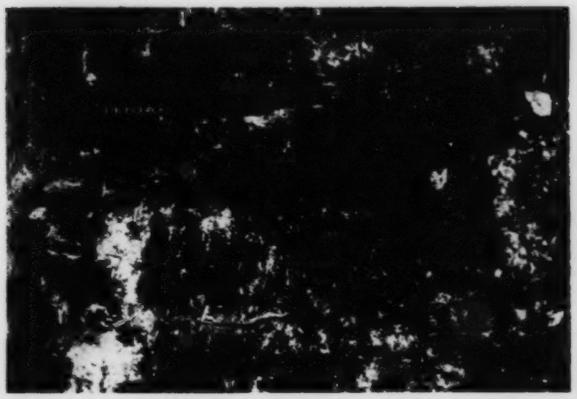


Plate 1. *Massive layered peridotite containing a band of disseminated chromite. Aggregate quarry, Route 360, 2 km south of South American Pond; lens cap in all plates is 5 cm in diameter.*

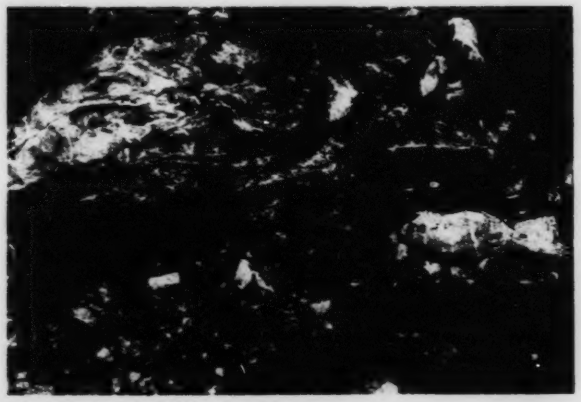


Plate 2. *Magnesite schist containing a chromite layer. Magnesite quarry, 0.75 km east of Route 360, 2 km south of South American Pond.*

Plate 3. *Mafic chlorite schist (Unit 4) from the outer margin of the amphibolitic schist that may form the basal aureole to the Great Bend Complex, 2 km northeast of South American Pond.*

indicated by the change in colour from black to cream, and by the presence of unaltered peridotite fragments in the sheared magnesite. Clearly the magnesite is derived from the adjacent peridotite. Shear zones also occur within the peridotite where the rock is a serpentine schist.

Banding is uncommon in the peridotite and where present is of limited extent. At the roadside quarry, 2.5 km southwest of South American Pond, 10- to 20-cm-thick, rhythmic bands are developed in the dunite with slightly chromite-rich dunite alternating with chromite-poor dunite (Plate 1). Locally the peridotite contains a pyroxene-bearing phase that is apparently harzburgite (Zwicker and Strong, 1986). The pyroxene has one well-developed cleavage and is probably enstatite. Banding is best developed 2 km west of Swan Lake. Here, a fault-bound block of Unit 1 within Unit 3 contains 15- to 50-cm-thick alternating layers of fine-grained dunite and harzburgite containing prominent pyroxene crystals. Overall, the banding is of limited extent and rarely extends from one outcrop to another.

Burton (1987) and Mercer (1988b) reported that massive and sheared serpentinite occur in drill core from the Chiouk Brook gold showing (Figures 1 and 2). The drilling intersected over 150 m of ultramafic rock and most holes ended in ultramafic rocks (see discussion in description of the Spruce Brook Formation).

Unit 2 has been previously mapped as a plug of pyroxenite that intruded the peridotite of Unit 1. Evidence for its intrusive nature is lacking as no contacts are exposed. It does, however, form a mappable unit as it contains a great variety of rocks in a very small area. The rocks include sheared, serpentinized dunite, banded, foliated, chromite-rich dunite and probable harzburgite, and massive, medium- and very coarse-grained, pyroxene-rich peridotite (possibly lherzolite). The western and northern boundaries of the unit are highly sheared. The southern and eastern boundaries are not exposed and are interpreted to be faults.

The southern portion of the Complex is dominated by sheared, magnesitized ultramafic rock (Unit 3, Figure 2; Plate 2). As discussed above, the magnesite is clearly an alteration product of Unit 1. The alteration is nearly complete and only locally is serpentinite seen within the magnesite. The magnesite is generally highly sheared and close to contacts with the adjacent units and the foliation is subparallel to the contact. An isolated occurrence of magnesite is found along the northwestern contact of Unit 1, 0.5 km west of South American Pond. Partially magnesitized serpentinite was also found near the contact with Unit 4, on the main woods road 1.5 km northeast of South American Pond. Quartz veins and disseminated quartz pods are a characteristic feature of parts of the magnesite unit.

Disseminated chromite and chromite layers occur throughout the magnesite. The best known example of chromite layering occurs within the NALCO quarry just east of the Bay d'Espoir Highway, 2 km west of Swan Lake (Figure 2). Here, the chromite forms a layer nearly 2 m long and 15 cm wide, dipping to the west at about 15° (Plate 2).

Approximately 300 m south of Lizard Pond, there is extensive, fine-grained silicification of the magnesite and widespread disseminated sulphide (see Mercer, 1988b; Evans, *this volume*). This general area has been extensively drilled and trenched to assess its gold potential.

Unit 4 forms a wide zone of intensely deformed mafic and ultramafic rocks along the northern margin of the Complex (Figure 2). The mafic rocks are converted to banded amphibolite or chlorite schist (Plate 3). The parental rock of the amphibolite, where it contains abundant granulated plagioclase, is probably of gabbroic origin. This is particularly well exposed along a disused woods road, 1.25 km due north of South American Pond. A series of nearly continuous outcrops exposes, from north to south, fine- to medium-grained, discontinuously banded amphibolite, medium- to coarse-grained plagioclase–quartz–(sericite) amphibolite, chlorite schist, and schistose to banded serpentinite. Regionally, the amphibolite outcrops are closer to the ultramafic rocks than the chlorite schist. Banding is locally well developed in the chlorite schist and the rock has the appearance of laminated, chloritized siltstone and the parental rock is possibly volcanic. The protolith of the fine-grained amphibolites is unknown.

Lineations are not well developed in the amphibolites or mylonites. The foliations are moderately to steeply-dipping to the south or southeast. In several places, folding of the main fabric is apparent (Plate 3) and in the main belt of Unit 4, chlorite schist contains folded second folds. At one locality, a possible C–S fabric was observed and this indicates sinistral shear.

The outcrops of amphibolite and chlorite schist that may have been observed by previous workers were apparently included with the sedimentary rocks of the Botwood Group, the Spruce Brook Formation, or an unnamed unit. The reassignment of outcrops on the north shore of the Northwest Gander River is also the first published indication that the Complex occurs north of the river.

A narrow belt of massive and cleaved, chloritized and epidotized, mafic tuff and basaltic flows (Unit 5; Figure 2) is found along recently constructed woods roads north of the Northwest Gander River. The volcanic rocks lie at the edge of the Spruce Brook Formation but are interpreted to be in tectonic contact with the formation as volcanic rocks are unknown elsewhere in the formation. The tuff contains feldspar fragments in a chloritized, cleaved matrix. Layers and laminations are well preserved in the more massive tuffs. Toward the eastern end of the unit, the rocks are possibly flows as the rocks are more thickly layered and generally lack features indicative of tuffs. Cleavage is also poorly developed.

Along a woods road south of the unit, a small outcrop of grey-green sandstone separates the mafic volcanic rocks from the amphibolites (Unit 4). This sandstone is probably part of the Spruce Brook Formation and is interpreted to form a fault-bound slice within the Complex. The proximity of the volcanic rocks to the Great Bend Complex possibly indicates that they may represent the upper portion of an ophiolite suite. Confirmation that the volcanic rocks are part of an ophiolite suite may be demonstrated by geochemical analyses in progress.

Along the southern margin of the Complex, outcrops of red, hematized and silicified rock (Unit 6) are well exposed in a trench at Breccia Pond. This rock is interpreted to be part of the Complex although Zwicker and Strong (1986) indicate that the rock is an altered sediment. Similar rocks occur in core obtained from the Lizard Pond area (D. Evans, personal communication, 1992). Geochemical analyses should give enhanced Ni and Cr values if the breccia is an altered ultramafic rock.

SPRUCE BROOK FORMATION–UNIT 7

The sedimentary rocks, which constitute the Spruce Brook Formation (Unit 7), were extended into the Eastern Pond map area by Colman-Sadd (1982). Unnamed quartz-rich sedimentary rocks and associated grey siltstone and slate (now part of the Spruce Brook Formation) were distinguished from greywacke and slate, conglomerate and volcanic rocks of the North Steady Pond Formation of the Ordovician Baie d'Espoir Group in the adjacent Burnt Hill map area (Colman-Sadd, 1985). Anderson and Williams (1970) had included these sedimentary rocks in the Silurian Botwood Group. Dec and Colman-Sadd (1990) reported that late Llanvirn to early Llandeilo fossiliferous limestone conglomerate unconformably overlie the Spruce Brook Formation and in combination with the regional association north of Gander Lake and in New Brunswick they suggested an age of Arenig or older for the formation.

The Spruce Brook Formation comprises interbedded, thick-bedded, grey and white sandstone and siltstone, and black slate (Unit 7; Plate 4); sandstone and minor siltstone containing abundant quartz veins constitutes subunit 7a (Figure 1). Parallel laminations are the most common sedimentary structure but locally convolute laminations and graded bedding may be found. Black slate outcrops 500 m north of the Northwest Gander River.

The area underlain by subunit 7a has been added to the Spruce Brook Formation (cf. Colman-Sadd, 1982). Its northern boundary is marked by a series of thick quartz and vuggy quartz–feldspar veins that may have been emplaced along a fault. The trace of this contact to the west would intersect with a faulted melange of mafic volcanic rocks in the southern part of Miguels Lake (see Colman-Sadd and Russell, 1988).

Subunit 7a consists of a monotonous sequence of grey-green, locally faintly red-weathering, thick-bedded, quartz-rich sandstone and minor siltstone. The unit is weakly cleaved parallel to bedding and contains interstitial, fine-grained chlorite. In comparison, the adjacent Botwood Group (Unit 14) contains a greater variety of rocks including red, green, grey and purple polymict conglomerate, and interbedded

Plate 4. *Complexly folded, grey sandstone of the Spruce Brook Formation. Great Gull River road, 1.5 km northwest of South American Pond.*

pebbly sandstone, sandstone and siltstone. Unit 14 contains a strong cleavage at a high angle to bedding. Thus, subunit 7a is petrographically and structurally distinct from Unit 14.

The Spruce Brook Formation is highly deformed with good examples of refolded folds and intersecting cleavages (Plate 4). West of Chiouk Brook, tight second folds of bedding and cleavage plunge gently to the east or west. This may indicate that a third generation of folds is present with a northerly trend. These sediments are virtually unmetamorphosed. Near the contact with the amphibolites (Unit 4) the sediments contain a fabric defined by chlorite. East of Chiouk Brook, a sequence of well-exposed psammites and semipelites are tightly folded into overturned folds that plunge to the southeast. The sequence is contact metamorphosed by the Mount Peyton Intrusive Suite with the progressive development of biotite, cordierite, and cordierite+garnet toward the contact.

At the Chiouk Brook gold showing, all published work (including Dickson, 1991) indicates that the area is underlain by sedimentary rocks. The black slate on Chiouk Brook is of particular significance as it contains significant pyrite–arsenopyrite mineralization and associated highly anomalous gold values (Zwicker and Strong, 1986; see also mineralization section). However, Burton (1987) and Mercer (1988b) reported that drilling in the vicinity of the Chiouk Brook gold showing intersected various types of ultramafic rock including serpentinite, magnesite, ultramafic schist and diabase, and also great thicknesses of graphitic breccia (subunit 1a; Figure 2). The latter is possibly equivalent to the graphitic, sulphide-bearing slate exposed in the brook. There are no exposures of serpentinite in the area but several angular glacial erratics were located north of the showing.

It is also possible that the sedimentary rocks exposed at the showing are not part of the Spruce Brook Formation but may form part of a melange zone that overlies the till-covered continuation of the Great Bend Complex. The strongly foliated rocks located to the north of the showing may not simply be part of the Spruce Brook Formation; they could represent the tectonic equivalent of the amphibolitic aureole (Unit 4) exposed 2 km to the southwest, along the Northwest Gander River (Figure 2).

BEDDED FOSSILIFEROUS CHERT–UNIT 8

Unit 8 occurs 1 km west of Bear Brook and also 4.7 km southeast of Webber Pond (Figure 1) where it is exposed in woods roads. These two widely separated areas are regionally along strike from each other. The unit is characterized by black chert that forms 1- to 2-cm-thick beds that are separated by 1- to 5-mm-thick laminae of graphitic, black shale. The Bear Brook occurrence consists of two outcrops of which the northern outcrop is a large disused road aggregate quarry. This occurrence was included in the Botwood Group by Zwicker and Strong (1986). However, at the Bear Brook occurrence, graptolites were found in the slate laminae and these have been identified by S.H. Williams of Memorial University. These indicate that the unit is late Caradocian in age. This age would indicate that the chert and slate are too young to be an integral part of the Great Bend Complex. Williams *(op. cit.)* suggests that the unit is probably equivalent to the upper part of the Lawrence Harbour Formation (*D. clingani* zone) (S.H. Williams, written communication, 1992).

DAVIDSVILLE GROUP–UNIT 9

The Davidsville Group occurs in the southeastern portion of the map area and also 5 km east of Webber Pond (Figure 1), where graptolite-bearing slates are correlated with the group. In the southeastern area, the Davidsville Group is assumed to be in stratigraphic contact with the Botwood Group along the valley southeast of Eastern Pond. Tallman (1989, 1990, 1991) indicates that the contact is a reverse fault that places the Botwood Group over the Davidsville Group. This fault coincides with a geophysical anomaly and also with a fault exposed in trenches at the Aztec showing, 6 km north-northeast of Eastern Pond (see Tallman, 1991).

The dominant rock types, in the southeastern area, are medium- to thick-bedded, grey to black slate and siltstone and grey to buff-coloured sandstone. Granule conglomerate and pebbly sandstone are minor components (Plate 5). The finer grained sedimentary rocks are commonly parallel laminated and locally contain crosslaminations. At one locality, a possible partial Bouma sequence was found that consisted of a lower, graded unit overlain by a massive central unit and a parallel laminated upper unit. The granule conglomerate and pebbly sandstone are poorly graded, and contain 2- to 4-mm-long, fragments of black chert, fine-grained felsic volcanic and abundant detrital feldspar (Plate 5). The chert fragments are similar to the chert in Unit 8.

East of Webber Pond, a 1-km-thick section of grey slates is exposed along the woods road. The slates are interpreted to be in fault contact with Unit 29 (Figure 1) although an unconformity is also a possibility. The contact coincides with a stream and there is an abrupt change in lithology. The sediments are also highly jointed, which may reflect late faulting. The nature of the western contact with the Botwood

Plate 5. *Coarse sandstone of the Davidsville Group containing chert, feldspar and volcanic clasts (Unit 9). Great Gull River road, 6 km east of Great Gull River.*

Group (Unit 10) is uncertain but it is assumed to be stratigraphic.

The slates east of Webber Pond contain a strong cleavage that is parallel to bedding. Certain horizons in the slate contain abundant graptolites. The graptolites have been examined by S.H. Williams and they indicate that the unit is late Caradocian to early Ashgill in age (S.H. Williams, written communication, 1992). This occurrence is probably a correlative of similar slates located 3 km to the northeast, on Beaver Brook, which contain the Caradocian graptolites *Climacograptus bicornis* (Hall), or *spiniferous* (Ruedemann), *Hallograptus?, Orthograptus?* and a stipe of *Dicellograptus* (J. Riva, written communication, 1980, in Blackwood, 1981). These strata have been included in the Davidsville Group by Blackwood (1981).

The Davidsville Group in the map area is not significantly metamorphosed. A prominent feature of the southeastern area of the Davidsville Group is the presence of two intersecting cleavages. The first cleavage is parallel to bedding and, where not affected by the second folds, is steeply dipping to the northeast. The second cleavage is axial planar to small, gently plunging, subhorizontal, second folds. The direction of plunge of the second folds is highly variable, indicating that they are also folded about open third folds. Quartz veins are locally concentrated along the axes of the second folds.

BOTWOOD GROUP (UNITS 10 TO 14) AND NORTH STEADY POND FORMATION (UNITS 10 AND 12)

Introduction

Units 10 to 14 have been assigned to the Botwood Group following the work of Anderson and Williams (1970), Blackwood (1981) and Colman-Sadd and Russell (1988). Units 10 to 12 lie east of the Mount Peyton Intrusive Suite and are generally continuous with units mapped by Blackwood (1981). Units 13 and 14 lie west of the Mount Peyton Intrusive Suite and form the southern termination of units mapped by Colman-Sadd and Russell (1988).

The Botwood Group has until recently been considered to be entirely of Silurian age and constitute a sequence of dominantly subaerial sedimentary and mafic and felsic volcanic rocks and minor, shallow-marine strata. Recent structural and stratigraphic analyses in the Bay of Exploits area indicate that the Botwood Group may contain a variety of marine sedimentary and mafic volcanic rocks (O'Brien, 1991). Moreover, recent fossil discoveries in turbidites of the Botwood Group have extended its range into the Upper Ordovician (Boyce *et al.*, 1991). In the Eastern Pond map area, rocks similar to those described by O'Brien (1991) are exposed in the central part of the map area although volcanic rocks are absent. The increased age range and variety of rocks may eventually require a revision of the Botwood Group.

Several areas have been removed from the Botwood Group where the underlying sedimentary rocks are poorly bedded, uncleaved, and poorly lithified; Units 29 and 30; cf. Anderson and Williams (1970), and Zwicker and Strong (1986).

The North Steady Pond Formation has normally been included in the Baie d'Espoir Group (e.g., Dec and Colman-Sadd, 1990). This has been based on an interpreted gradational contact of the formation with other units of the group. The North Steady Pond Formation (part of Units 10 and 12), outcrops in the southwestern corner of the map area, in the vicinity of Swiss Lake, where it forms the continuation of the strata in map areas NTS 2D/5 (Colman-Sadd, 1985) and 2D/12 (Colman-Sadd and Russell, 1988).

To the northeast of Swiss Lake, the formation appears to be continuous with and contain the same rocks as the various units of the Botwood Group. However, stronger support for the correlation of the North Steady Pond Formation with some units of the Botwood Group comes from a new fossil locality on the Bay d'Espoir Highway (Figure 2). At this locality, one bed of calcareous sandstone contains abundant fragments of crinoid stems, one brachiopod fragment, and bryozoa (W.D. Boyce, personal communication, 1991). This locality contains a fauna similar to that described by Anderson and Williams (1970) and Blackwood (1981). They reported that on Cooper Brook (Figure 1), a calcareous siltstone contains corals, crinoid stems and brachiopods interpreted to be indicative of a Silurian age.

Unit Descriptions

Unit 10 is well exposed along the Great Gull River and along the eastern part of the Northwest Gander River. Sporadic exposures occur along and near the woods roads east of the Great Gull River. The eastern part of the unit comprises a sequence of northwest-dipping, medium- to thick-bedded grey, brown-weathering sandstone and coarse-grained sandstone, white quartz-rich sandstone and minor, thin-bedded, grey-green and purple siltstone. Two areas of concentrated black slate float are exposed along the main branch road. The source of the float is not apparent. Slump structures, convolute lamination, graded bedding, crossbedding, parallel and crosslaminations all occur in these sediments. Evidence for shallow-water sedimentation is not apparent and these sediments are interpreted to be possibly proximal turbidites. The sedimentary features generally indicate younging to the northwest.

Along Great Gull River, the most southerly exposures of Unit 10 comprise a sequence of relatively schistose, sericitic and chloritic sedimentary rocks with numerous small folds. Immediately to the northwest, a monoclinal, northwest-dipping sequence of fine- to medium-grained sandstone and minor siltstone and slate are exposed nearly to the mouth of the river. These strata are much less deformed and sedimentary structures are well preserved. The sandstone is medium- to thick-bedded, grey-green and green and rarely brown.

The most conspicuous sedimentary structures are mudcracks and ripple marks. The mudcrack polygons are on average about 20 cm in diameter but they are commonly stretched. The length to width ratio is 2:1, in a few places, with the long axis plunging to the north. The ripple marks are mainly symmetrical and have wavelengths of 20 to 40 cm and amplitudes of 2 to 3 cm. The rare asymmetrical ripples (Plate 6) indicate current flow towards 063 and 276°. Other sedimentary features include parallel laminations, sand lenses, disrupted laminations, crosslaminations, scours and graded beds.

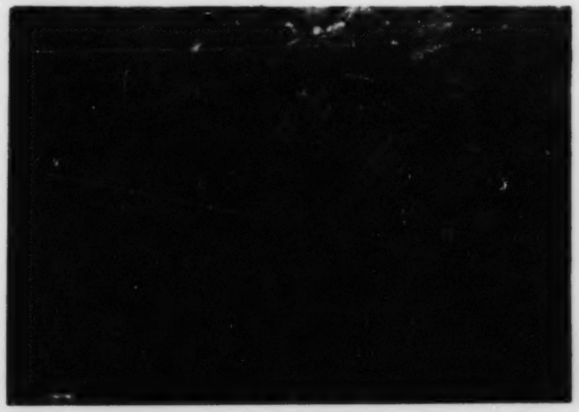


Plate 6. *Asymmetric ripple marks in sandstone of the Botwood Group (Unit 10). Great Gull River, 3 km south of the mouth of the river.*

In the vicinity of Miguels and Rolling falls, Unit 11 (the hornfels equivalent of Unit 10) is a sequence of medium- to thick-bedded sandstone that contains mudcracks, parallel- and crosslaminations, graded beds, scours, and siltstone rip-ups. The effects of contact metamorphism by the adjacent gabbro (Unit 17) are reflected in the grey rather than buff or green colour of the sandstone, and toughening of the rock up to 2 km from the contact. At the contact at Miguels Falls, the rock is partially melted for about 1 m from the contact. The succession at Miguels and Rolling falls is similar to that along Great Gull River with medium- to thickly-bedded sandstone that commonly display sedimentary features including abundant mudcracks.

Near the mouth of Bear Brook, a well-exposed sequence of thin-bedded, hard, green sandstone and minor siltstone dips steeply to the east. Elongated, sandy limestone nodules form strings parallel to bedding at one locality. The sequence generally lacks sedimentary structures but the thin, parallel beds, commonly 5 to 20 cm thick having thin silty partings are a prominent feature. This sequence is almost continuously exposed for 300 m upstream to where it is probably unconformably overlain by soft, massive, poorly bedded, coarse-grained sandstone and red and green, thin-bedded shale (Unit 29). Three kilometres upstream, a similar sequence of thin-bedded sandstone is exposed.

Along the Northwest Gander River, 1.7 km east of Great Gull River, a sequence of green, thin- to medium-bedded, parallel-bedded sandstone, siltstone and slate is nearly continuously exposed downstream for 4 km. The sediments only rarely display sedimentary features such as graded beds and crosslaminations. These features indicate that the sequence is consistently younging to the west.

One kilometre south of Jumbo Brook (Figure 1), grey-green thin to medium-bedded sandstone is in faulted contact with very thick-bedded pebble conglomerate and sandstone units (subunit 10a) that are several metres thick. The thick-bedded sandstone and conglomerate structurally underlie the thinner bedded sandstones. Across the fault, there is an abrupt change in the dip of bedding. The thin beds dip gently to the west at 30 to 45° whereas the conglomerate dips to the east at 70°. This fault may represent the trace of tight overturned folds and thrust structures. The strata exposed discontinuously downstream for about 6 km consistently dip to the east at high angles. Sedimentary features such as grading, crossbedding and scours in the thick-bedded sandstones and pebble conglomerates and scours in the thin-bedded sandstones indicate that the sequence is overturned to the west. The thick-bedded strata form lenses within the thin-bedded strata.

The thin-bedded strata comprise a sequence of grey, parallel-bedded, sandstone and minor siltstone that forms long outcrops of continuous beds. Locally, ovoid nodules of sandy limestone form bands in the thin-bedded sandstones. The nodules and the thin, parallel beds are very similar to those on Bear Brook. A similar conglomerate sequence is exposed at the southwestern end of the belt around Swiss Lake. The thick-bedded sandstones and conglomerates on the Northwest Gander River (subunit 10a) contain clasts of black chert, siltstone and red jasper. The chert fragments resemble the bedded chert exposed southeast of Webber Pond and the jasper is presumably derived from the Gander River ophiolite complex to the east (see Blackwood, 1982). At Swiss Lake, the conglomerate contains abundant felsic volcanic clasts along with conspicuous jasper. The volcanic fragments are probably derived from the North Steady Pond Formation southwest of the map area (Colman-Sadd, 1985).

Southeast of Webber Pond, the grey sandstone and siltstone along the contact with the Mount Peyton Intrusive Suite are tentatively included in Unit 10 of the Botwood Group. The unit here consists of medium-bedded grey sandstone and slate that generally lack sedimentary structures but a few ripple-marked surfaces are present. Outside of the contact aureole, the sandstone and slate are strongly cleaved.

The sedimentary features of Unit 10 indicate that much of the lower part of the sequence was deposited as possibly proximal turbidites. The thick sandstone and conglomerate beds (subunit 10a) may represent debris flows. The turbidites were then overlain by progressively shallower water deposits in very shallow-marine conditions and it is probably an intertidal sequence from Great Gull River to Rolling Falls.

The Botwood Group strata along the lower portion of the Northwest Gander River requires further study. In particular, the relationship of the turbiditic sandstone and associated conglomerates to the green slates and shallow-water strata is uncertain. Blackwood (1982) describes thin-bedded sandstones of the Davidsville Group (see Plate 12) that are virtually identical to the thin-bedded sandstones exposed along the Northwest Gander River and considered to be part of the Botwood Group.

A feature of Units 10 and 11 is the presence of a variably developed cleavage parallel to bedding. Variations in direction of dip along the Northwest Gander River indicate that the cleavage and bedding are folded into open folds that plunge gently to the north. This plunge coincides with small folds along the Northwest Gander River, 4 km northeast of Great Gull River, that plunge toward 342° at 16°. As cleavage is also folded by these folds, the main cleavage must reflect a recumbent first fold and the open folds are second folds. The presence of beds that are overturned to the west may be interpreted to indicate that the first fold is a recumbent syncline. Locally, folds of cleavage plunge gently to the southwest or moderately to the west and this may reflect a third phase of folding that forms open folds that mainly plunge to the east at 40 to 70°. A cleavage is rarely associated with the second folds.

Regional metamorphism of Unit 10 is in the lower-greenschist facies with the widespread occurrence of chlorite reflected in the green colour of the rocks. Contact metamorphism has converted the strata to hornfels near the contact with the Mount Peyton Intrusive Suite and small cordierite porphyroblasts are found in the sandstones 6 km east of Webber Pond.

Unit 12 comprises a thin- to medium-bedded sequence of red and minor grey and beige siltstone and sandstone that appears to lie within Unit 10. The unit is a steeply dipping to overturned sequence that is well-exposed along woods roads and adjacent cut-over areas west of Spud Pond. This area of Unit 12 is a fault-bound block that is surrounded by components of the Great Bend Complex; the contacts are not exposed. To the northeast of Spud Pond, the unit youngs toward the sandstone of Unit 11. However, there is a change in the orientation of bedding that possibly indicates that there is a faulted contact.

The sedimentary features of Unit 12 include crosslaminations, mudcracks, graded beds, shale rip-ups and rain pits. These clearly indicate very shallow water and, by association with the marine Unit 10, possibly intertidal conditions. The unit is unmetamorphosed and west of Spud Pond contains a poorly developed, east-trending cleavage that is at an angle to bedding. The intersection of these planes indicate a fold that plunges due north at 80°.

West of the Mount Peyton Intrusive Suite, a sequence of buff-weathering, grey sandstone hornfels (Unit 13) forms part of the Botwood Group. The areal distribution of the unit is quite different from that shown by previous workers as it actually forms a large east-trending embayment into the intrusive suite (cf. Anderson and Williams, 1970).

Unit 13 consists of medium-bedded, parallel laminated sandstone locally containing abundant mudcracks, and minor slate and felsic volcanic rocks. The sandstones are extremely hard and this is interpreted to be the result of widespread contact metamorphism by the Mount Peyton Intrusive Suite. At the northern end of the unit, where the sandstone is in contact with granite (Unit 21), the sandstone is metamorphosed in a much narrower zone. However, there is clear evidence of partial melting of the sandstone. Small veins and patches of tourmaline–biotite–muscovite granite and pegmatite occur in the sandstone (subunit 13a) within 300 m of the contact. Abundant xenoliths of partially melted and/or granite-veined sandstone define the northern margin of Unit 13. Within the adjacent granite, sandstone xenoliths are locally abundant.

Extensive subcrop of felsic volcanic rocks occur along a branch woods road, 2.2 km northeast of Rolling Pond. The rock is red, fine-grained, massive porphyritic rhyolite. One km northeast of Rolling Pond, crystal-lithic tuff is exposed at the end of a woods road. This rock is, however, similar to the glacial erratics of tuff that occur throughout the map area and the occurrence may be a partially buried erratic.

The beds in Unit 13 are generally parallel but have been boudinaged, possibly tectonically, as well as by soft-sediment processes. The minor slates possess well-developed cleavage and a spaced cleavage is locally present in the sandstone beds.

Unit 13 is similar to Unit 11 and the western portion of Unit 10 and is probably a correlative. All are dominated by grey, medium-bedded sandstone, contain mudcracks and exhibits various types of lamination.

Unit 14 comprises a highly variable sequence of well-cleaved, red, purple or grey-green conglomerate, sandstone and siltstone in strong contrast to Unit 13. The change is abrupt although not exposed. There is also no indication of contact or regional metamorphism. The stratigraphically lowest part of the unit in the map area is a poorly sorted, pebble and cobble conglomerate located near the contact with subunit 7a. The clasts consist mainly of grey sandstone, quartz-veined green sandstone, and of quartz-rich sandstone. The beds vary in thickness from 1 to 3 m and are probably lenses.

The sandstone is also polymict and displays parallel, convolute and crosslaminations, scours, rip-ups, and sandstone dykes. The siltstone includes purple, green, red and grey varieties that contain the same types of laminations as the sandstone. Beds vary in thickness from 1 cm to 1 m. The sequence probably represents fluvial deposits.

A prominent feature of the siltstone and sandstone is the strong cleavage at a high angle to bedding (Plate 7). Minor folds of bedding plunge steeply to the east-southeast and contain an axial-planar cleavage.

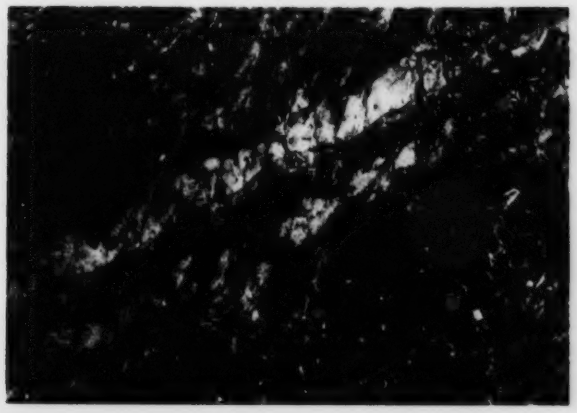


Plate 7. *Strongly cleaved, thin-bedded sandstone of the Botwood Group (Unit 14), 2 km northwest of Rolling Pond.*

The distinctly different rocks of Units 13 and 14 and other factors such as the extensive contact metamorphism of Unit 13 and the strong cleavage in Unit 14 may indicate that there is an unconformable relationship between them. Unit 13 was possibly metamorphosed by the intrusive suite before deposition of Unit 14. The deformation that produced the cleavage in Unit 14 may not have been of sufficient intensity to produce a penetrative fabric in the hornfels of Unit 13.

POST-KINEMATIC SEDIMENTARY ROCKS

Four areas of post-kinematic sedimentary rocks have been defined in the southeastern part of the map area. These rocks lack any penetrative fabric, metamorphism or minor intrusions and a Devonian or younger age is assumed.

Unit 28 forms a prominent hill, 2.5 km west of Spud Pond (see also Colman-Sadd, 1982) that is underlain by poorly bedded, sand- to boulder-size clasts. The largest clast found is about 5 m long. The clasts are composed entirely of ultramafic rocks and the dominant variety is peridotite (Plate 8). The peridotite includes layered and massive types, and also asbestos-bearing and quartz-veined clasts. Colman-Sadd (1982) reported the presence of magnesite clasts. Bedding is defined by poorly developed, gradual changes in clast size over a thickness of 2 m.

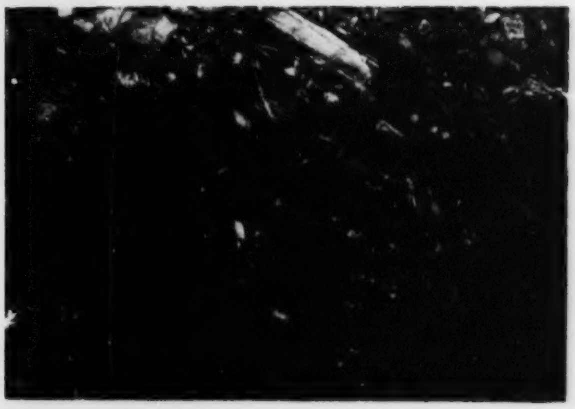


Plate 8. *Peridotite-cobble conglomerate of Unit 28. Small isolated hill, 0.5 km north of Lizard Pond.*

The conglomerate overlies schistose peridotite which outcrops immediately to the north of the hill, and sheared magnesite that lies to the south of the hill. These two ultramafic units are in sheared contact. The ultramafic clasts are derived from massive peridotite and the nearest exposed source is 1 km to the west, near South American Pond.

Unit 29 occurs in two extensive areas. The main area lies west of the Great Gull River and the other smaller but potentially economically significant unit is located to the north, 7 km east of Webber Pond (Figure 1). The dominant rock type in the southern area is poorly sorted feldspathic sandstone accompanied by minor red and black shale, pebble and cobble conglomerate. Bedding is extremely poorly developed and this is a characteristic feature of the unit. The sandstone unconformably overlies Unit 10 on Bear Brook and pebble conglomerate nonconformably overlies a diabase dyke (Unit 27) 300 m east of Bear Brook (see Dickson, 1991). The contacts with the other adjacent units are unexposed.

The cobble conglomerates outcrop 200 to 500 m west of the mouth of Great Gull River. The main clast type is grey sandstone (similar to the sandstone hornfels of Unit 11 of the Botwood Group) pink felsic volcanic clasts, chert and quartzite. The matrix is rich in feldspar and small sandstone clasts. The pebble conglomerate contains conspicuous red jasper and black chert fragments, along with beige and pink felsic volcanic fragments, quartz-rich sandstone, quartz-veined sandstone, and fine-grained pink granite. The black chert is similar to Unit 8; the felsic volcanic rocks have presumably been derived from the North Steady Pond Formation, west of the map area (Colman-Sadd, 1985); the jasper is possibly derived from an ophiolite complex (see e.g., Blackwood, 1982). The quartz-rich and quartz-veined sandstone fragments are probably derived from the Spruce Brook Formation.

The cobble conglomerate is assumed to be in contact with red shale and this is in assumed contact with black shale. No shale bedrock is exposed but thick talus slopes of shale are found along the woods road.

The sandstone is dominated by 2- to 4-mm-long plagioclase feldspar fragments and mafic mineral detritus. The proportions are similar to that found in the nearby gabbro intrusions (Units 15, 17 and 18), which are presumed to be the main source of sand-sized clasts.

A rather unusual unit is exposed along the Northwest Gander River for 1.5 km east of Great Gull River. The unit is a pebble to cobble conglomerate having well-aligned larger clasts set in a coarse sand- to granule-conglomerate matrix. The unit is poorly lithified and is easily fragmented. However, the conglomerate has sufficient competence to be cut by a fault that is filled with a quartz vein. This also indicates that the unit is probably not Pleistocene.

The unit contains a wide variety of clasts; these include grey, green and red sandstone, quartz-veined sandstone, grey coarse-grained sandstone, red siltstone, green slate, and amphibolite schist. The slate fragments are angular and the other clasts are rounded and elongate. All these rocks can be matched with the nearby units. The flatter, large fragments are aligned and this is assumed to represent bedding. However, it is also possible that the fragments are imbricated and the bedding planes dip at shallower angles. One 30 cm boulder is surrounded by flat clasts that are wrapped around its structurally lower margin. The unit bears a strong resemblance to the overlying Pleistocene glaciofluvial gravels and a fluvial source is likely.

MOUNT PEYTON INTRUSIVE SUITE AND ASSOCIATED INTRUSIONS (UNITS 15 TO 27)

Units 15 and 16 are gabbro intrusions that are located at Bear Pond and along the Northwest Gander River, respectively. Most earlier workers assigned these units to the Great Bend Complex mainly because of their spatial relationship to the ultramafic rocks. Zwicker and Strong (1986) indicated that the gabbro near Bear Pond was spatially separated from the ultramafic rocks by sedimentary rocks and, therefore, unrelated to the ultramafic rocks. Recent woods-road construction, however, has exposed several gabbro outcrops that indicate the gabbro is in contact with the ultramafic rocks. The gabbro near the contact is extensively brecciated indicating that the contact is faulted. The contact of the gabbro with the Botwood Group is probably intrusive as diabase dykes cut the Botwood Group, northeast of Bear Pond. Widespread unmetamorphosed, red siltstone float overlies the gabbro and is possibly related to Unit 29.

The contact of the gabbro with the ultramafic rocks is assumed to be a fault but E. Çogulu (Geological Survey of Canada, personal communication, 1991) has suggested that the adjacent ultramafic rocks are contact metamorphosed indicating an intrusive contact. It is unlikely that the gabbros are related to the Great Bend Complex, as similar rocks occur within the Mount Peyton Intrusive Suite, but chemical analyses should resolve this.

The gabbro of Unit 15 ranges in grain size and texture from medium-grained, equigranular diabase and very coarse-grained gabbro, with a weak alignment of pyroxene and plagioclase, to very coarse-grained massive gabbro. The very coarse-grained gabbro contains 2-cm-long plagioclase crystals and 1-cm² interstitial hornblende.

Unit 16 gabbro is dominantly medium- to coarse-grained and equigranular but at one locality, on the south side of the Northwest Gander River, three outcrops of the gabbro are layered. The eastern contact with the Botwood Group is faulted and the gabbro is weakly foliated and brecciated for about 100 m west of the contact. The remainder of the gabbro is undeformed. The layered gabbro contains parallel and continuous bands that have varying proportions of plagioclase and pyroxene (Plate 9); some layers are pegmatitic. The bands vary in thickness from 30 to 60 cm. They are also laminated and dip steeply to the southeast. Generally, the gabbro is quite massive forming large homogeneous blocks of white-weathering, grey gabbro.

Plate 9. *Layered, massive gabbro of Unit 16. Northwest Gander River, 2 km northwest of Spud Pond.*

THE MOUNT PEYTON INTRUSIVE SUITE

The Mount Peyton Intrusive Suite comprises 10 contiguous units of massive plutonic rocks in the Eastern Pond map area. Intrusive relationships indicate that the granitic rocks are younger than the gabbroic rocks and that the more leucocratic granites are the youngest intrusions. The age relationships of the various gabbroic units are uncertain due to lack of intrusive contacts.

Previous radiometric age determinations on the suite have been made to the north of the map area on samples from map areas NTS 2D/14 and 2E/3 (Table 1). The 270 ± 52 Ma

Table 1. Published radiometric ages for the Mount Peyton Intrusive Suite. Ages have been recalculated to current decay constants where necessary (from Mandville, 1989)

Rock type	Age (Ma)	Method	Location	Reference
Diorite	418 ± 21	K–Ar biotite	Burnt Lake	Wanless *et al.*, 1965
Granodiorite	270 ± 52	K–Ar hornblende	NW Gander R.	Wanless *et al.*, 1967
Granodiorite	369 ± 21	K–Ar biotite	Rattling Lake	Wanless *et al.*, 1967
Granite, gabbro	380 ± 30	Rb–Sr whole rock	N. end pluton	Bell *et al.*, 1977
	390 ± 15	Rb–Sr whole rock		revised by Bell and Blenkinsop, *in* Reynolds *et al.*, 1981
Granodiorite	428 ± 5	Ar–Ar biotite	Frozen Ocean L.	Reynolds *et al.*, 1981
Gabbro	423 ± 5	Ar–Ar hornblende	Trans-Canada	Reynolds *et al.*, 1981
Gabbro	417 ± 5	Ar–Ar biotite	Trans-Canada	Reynolds *et al.*, 1981
Gabbro	410 ± 5	Ar–Ar hornblende	Trans-Canada	Reynolds *et al.*, 1981
Gabbro	414 ± 5	Ar–Ar biotite	Trans-Canada	Reynolds *et al.*, 1981
Gabbro incl.	412 ± 8	Ar–Ar biotite	Trans-Canada	Reynolds *et al.*, 1981
Gabbro incl.	400 ± 7	Ar–Ar hornblende	Trans-Canada	Reynolds *et al.*, 1981
Granitic matrix	422 ± 8	Ar–Ar biotite	Trans-Canada	Reynolds *et al.*, 1981

(K–Ar hornblende; recalculated) date is considered to be suspect as the hornblende is apparently chloritized. Similarly, the 410 ± 5 Ma is suspect due to the replacement of hornblende by fibrous amphibole and sericite. Reynolds *et al.* (1981) note that the dated biotite is quite fresh and suggest an age of 420 Ma for the gabbro. They also report a revised age of 390 ± 15 Ma (K. Bell and J. Blenkinsop, personal communication, 1980) for the granite phase.

Preliminary U–Pb (zircon) work on pegmatitic hornblende gabbro (Plate 10) in massive, medium-grained gabbro (Unit 18) on Rolling Brook, 1 km south of Rolling Pond (dated under contract by G.R. Dunning, 1991) also indicate a Silurian age. When completed, this work will provide a precise age of part of the Mount Peyton Intrusive Suite. As the contact metamorphic aureole overprints the fabrics in the adjacent sedimentary units, including the Spruce Brook Formation and Units 10, 11 and 13 of the Botwood Group, all these units must be older than the intrusion and have been deformed before it.

Unit 17 is assumed to be the oldest unit of the suite in the map area. It is characterized by well-developed, parallel igneous layers and a generally coarse grain-size. The layering strikes northerly and is apparently truncated to the north by poorly layered to uniform, medium-grained gabbro (Unit 18; Plate 11). This is taken to indicate that Unit 17 is the older unit.

The eastern and western contacts of Unit 17 with the Botwood Group (Unit 11) are exposed on the Northwest Gander River at Rolling Falls and Miguels Falls. The contacts are fairly sharp with minor assimilation of the adjacent sandstone. The gabbro is slightly finer grained within 50 cm of the contact and medium-grained away from the contact; it also displays a swirly flow foliation. Ophitic intergrowths of plagioclase and hornblende occur in patches. Small, round xenoliths of sandstone are found in the gabbro up to 30 cm from the contact. The contacts are steeply dipping and at Miguels Falls the gabbro is concordant with, and appears to underlie, the sandstone layers. No contact metamorphic minerals were observed and the sandstone is a hornfels.

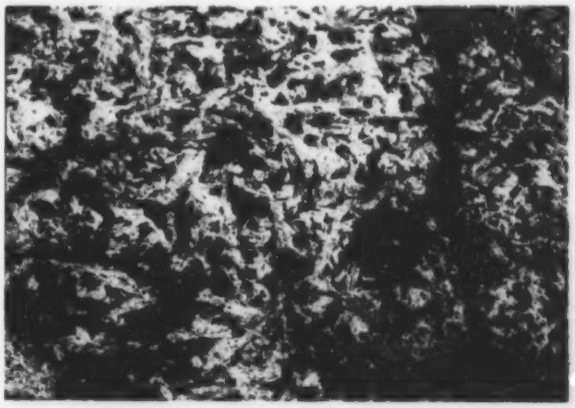


Plate 10. *Pegmatitic gabbro of the Mount Peyton Intrusive Suite (part of Unit 18). Rolling Brook, 1 km south of Rolling Pond.*

Plate 11. *Massive, medium-grained hornblende gabbro. Unit 18 of the Mount Peyton Intrusive Suite, 1 km east of Rolling Pond.*

The contact with the Spruce Brook Formation is exposed in the woods, 500 m north of the Chiouk Brook gold showing. Fine-grained massive diabase cuts strongly foliated, quartz-rich sandstone. There is also a metamorphic aureole about 1 km wide in the Spruce Brook Formation, 2.8 km south of Rolling Pond, which is defined by the progressive development of biotite, cordierite, and lenses containing garnets toward the contact. Within the gabbro, veins of possibly autobrecciated gabbro cut through the massive fine-grained gabbro and may indicate that the gabbro is a high-level intrusion.

The gabbro is dominated by plagioclase, pyroxene and locally conspicuous hornblende. In the layered units, the plagioclase : pyroxene ratio varies from 10:1 to 1:2. The layering is parallel and straight but the junction of the layers is generally ragged. The layers are 4 to 30 cm thick. Along Rolling Brook, the layers are nearly vertical and trend north-northwest but to the north of Rolling Falls the layers dip gently to the northwest. Thin veins and patches of magnetite occur in the gabbro 1.8 km west-northwest of Rolling Falls. Near the northwest contact of the unit, the gabbro outcrops are altered to mounds of extremely friable and rusty gossan.

Unit 18 is the most extensive unit of the suite in the map area occurring in three apparently separate areas. The texture is also highly variable and ranges from fine-grained, uniform gabbro having clots of amphibole (Plate 11) to coarse-grained, trough-layered gabbro. The eastern and northern portions are dominantly uniform gabbro and the layered gabbro is restricted to the Rolling Pond area.

Unit 19 is distinguished from Unit 18 only by the presence of abundant granite dykes, which form about 10 percent of Unit 19. The dykes are probably correlatives of the granodiorite and granite of Units 23 and 24, respectively, with which the gabbro is in contact. The contacts with the country rocks are not exposed but the extensive sandstone hornfels of Unit 13 and the narrow hornfels in Unit 10, east of Webber Pond, are almost certainly metamorphosed by Unit 18.

Generally, the gabbro is fine- to medium-grained pyroxene ± hornblende gabbro containing widely scattered inclusions of medium- to medium coarse-grained poikilitic gabbro. Locally, the gabbro contains conspicuous 3-mm-long hornblende needles that are strongly aligned. The layered gabbro contains highly irregular layers of medium, coarse-grained and pegmatitic gabbro. The pegmatitic gabbro contains abundant hornblende crystals locally over 10 cm long. The pegmatitic gabbro is in gradational contact with the finer grained gabbro and forms layers rather than veins. Pegmatite also forms patches of various sizes throughout the unit.

The dykes that cut Unit 19 are all granitic and occur throughout the unit. The proportion of dykes varies from 1 to 20 percent and they range in thickness from about 5 cm to 2 m. Individual dykes may vary in orientation and thickness. The dykes are white to grey, massive, leucocratic, fine- to medium-grained, biotite ± hornblende granite. Ovoid and angular gabbro xenoliths from Unit 19 are abundant in the adjacent granite (Unit 25), south of Otter Lake.

The gabbro of Unit 20 is probably cut by the granite of Unit 21; exposures of these units are in close proximity along woods roads 2.5 km southwest of Haynes Lake in map area NTS 2D/12 (see Colman-Sadd and Russell, 1988). Unit 20 is also in exposed intrusive contact with the Botwood Group along the woods roads in this area. The gabbro contains abundant sedimentary xenoliths and gabbro dykes cut the sandstone and is converted to hornfels.

The gabbro of Unit 20 is distinguished from that of Unit 18 by the presence of variable amounts of 1- to 2-mm diameter biotite flakes in an otherwise similar matrix. The gabbro contains medium-grained and slightly coarser-grained varieties as xenoliths and is cut by several granitic dykes and hornblende gabbro pegmatite veins. It contains recognizable hornblende in hand specimens but the main ferromagnesian mineral is probably pyroxene.

Unit 21 is a fine- to medium-grained, massive, leucocratic muscovite–biotite granite that occurs along the western margin of the suite. The actual extent of the unit is a matter of definition. This is due to the presence of locally extensive sandstone outcrops that are cut by abundant granite veins and also granite having abundant sandstone inclusions. Furthermore, many of the xenoliths are migmatized and contain sizeable patches of granite. The large protrusion of Unit 13 sandstone into the suite reflects the interpretation of a series of sandstone outcrops, some of which are veined by granite, as part of Unit 13. The outcrops along the northern portion of the protrusion were mapped as part of the granite by Colman-Sadd (1982) rather than as part of the Botwood Group.

The granite is in intrusive contact with Unit 13 south-southwest of Haynes Lake. Several granite and granite pegmatite veins cut the sediments adjacent to the contact and there is incipient migmatization. The neosome is similar to the main granite but also contains conspicuous tourmaline and fine-grained muscovite and biotite. It forms isolated patches in the sandstone and is clearly not derived directly from the main granite.

The formation of the migmatite by intrusion of the granite is possibly unlikely as the area underlain by the granite (Unit 21) is quite small and also quite narrow where migmatization has occurred. This volume of granite is unlikely to have contained sufficient heat to migmatize the country rocks. An alternative scenario is that both the granite of Unit 21 and the migmatization are related to the intrusion of the adjacent gabbro, which caused extensive partial melting of the adjacent sedimentary rocks. The gabbro is much more extensive, would have been emplaced at a much higher temperature, and therefore would have contained more heat. This also would explain the presence of extensive areas of sandstone xenoliths, migmatization and granite veining of the sandstone inclusions within the granite unit and adjacent to the gabbro. This scenario also would indicate that Unit 21 granite is the oldest granite in the map area.

Unit 22 consists of two small areas of grey, massive, fine-grained, biotite–hornblende tonalite. The smaller area is located along the contact of the suite with the sandstone of Unit 13. The contacts of both areas of tonalite with the adjacent gabbro (Unit 18) are not exposed but the tonalite is interpreted to be the younger unit as it is not cut by any mafic dykes and one area is surrounded by gabbro.

The granodiorite of Unit 23 is located to the north of Webber Pond where it forms part of an area dominated by granitic rocks. The unit is mainly a grey, medium-grained, massive, biotite–hornblende granodiorite that is locally coarser grained. The unit is cut by granite dykes from Unit 25 and is intruded by Unit 24. The contacts with Unit 24 are exposed southwest of Crowe Lake and northeast of Webber Pond. The contacts are sharp and veins of Unit 24 cut the granodiorite.

Unit 24 is a grey to pink, leucocratic, massive, fine-grained, biotite granite that is well exposed along the main road north of Webber Pond. Small, angular, mafic xenoliths occur locally in the unit. The unit appears to form a roughly circular plug that has intruded all adjacent units.

Unit 25 underlies an extensive area in the northeastern portion of the map area. This unit is shown by Blackwood (1981) as continuing to the northeast and forming the eastern portion of the intrusive suite. The contact of the granite with the adjacent sandstone of Unit 10 is not exposed but the sandstones are contact metamorphosed. A pink granite dyke (see Dickson, 1991) cuts Unit 10, 2.5 km southeast of Webber Pond and although the dyke is 4 km from Unit 25 it is probably related to it. The granite is in sharp contact with the gabbros of Units 18 and 19, northeast of Webber Pond. Adjacent to the contact with Unit 25, east of Webber Pond, the granodiorite (Unit 23) is cut by numerous pink granite dykes that are clearly derived from Unit 25.

Unit 25 is consistently a pink, massive, fine-grained, leucocratic, biotite granite (Plate 12) that locally contains miarolitic cavities. Biotite forms less than 1 percent of the rock. Blackwood (1981) notes that micrographic intergrowths are apparent in thin section. These features indicate that the granite is a high-level intrusion. Unit 25 is probably the youngest unit of the suite, at least in the map area.

Plate 12. *Massive, fine-grained granite of Unit 25 of the Mount Peyton Intrusive Suite. Woods road, 3.5 km south of Otter Lake.*

Unit 26 comprises all the diabase dykes found within the Mount Peyton Intrusive Suite. Several 10-cm-thick, gently dipping, light-grey, diabase dykes cut the gabbro (Unit 18) along Rolling Brook. These truncate layers and commonly have thin offshoots. Farther to the east, several 1-m-thick, dark-grey, medium-grained diabase dykes also cut the gabbro. These dykes have irregular margins and their trend is inconsistent.

Unit 27 comprises all the diabase dykes found outside the Mount Peyton Intrusive Suite. These dykes are all grey, medium-grained and massive. One cuts the Botwood Group sandstone hornfels (Unit 11), 2 km southwest of Rolling Falls. Several dykes cut the green slates and sandstones along Great Gull River and 2 km to the east and 1.5 km west of the river. A gently dipping dyke cuts the Spruce Brook Formation at the Chiouk Brook gold showing. Another dyke is interpreted to have intruded the ultramafic rocks along the Northwest Gander River (see Dickson, 1991 for locations of all these dykes).

SURFICIAL GEOLOGY

The glacial history and striation data for the map area are contained in reports by Taylor and St. Croix (1989), St. Croix and Taylor (1990). The data indicate an early ice flow to the south, with subsequent flows to the east and then north. Extensive glacial outwash and till deposits occur throughout the map area but sand and gravel outwash deposits are most extensive along the Northwest Gander River valley (this study). Glacial erratics are abundant and boulders of the Stony Lake Volcanics (Colman-Sadd and Russell, 1982) are particulary conspicuous and widespread. Some of these have travelled a minimum distance of 30 km.

MINERALIZATION

Mineral exploration of the area began following the report of magnesite by Grady (1953). The Great Bend area received a cursory examination by Harrison (1953) and was mapped in detail by Coleman (1954), both of the Newfoundland and Labrador Corporation Limited (NALCO). A bulk sample of magnesite was obtained for metallurgical testing (Newfoundland and Labrador Corporation Limited, 1953) and subsequently reported to contain pervasive fine-grained silica that presented a problem in obtaining high-quality magnesite (Newfoundland and Labrador Corporation Limited, 1954; Stewart, 1957).

In 1963, three magnesite showings and an outcrop of sheared serpentinite, in the vicinity of Lizard Pond, were drilled to depths of 17 to 87 feet (approximately 5 to 26 m; Bell Asbestos Mines Limited, 1963). It is here assumed that this was an assessment of the magnesite potential of the area as NALCO had never indicated that asbestos was significant in the Great Bend area.

Magnesite rock from the Great Bend area was petrographically analyzed by Soles (1972) for NALCO. Soles reported that fine-grained talc and variable amounts of quartz along with minor chromite, magnetite and pyrite occur with magnesite. He also concluded that beneficiation of the magnesite ore could be difficult.

More recent exploration in the map area has concentrated on the gold potential of the Great Bend Ultramafic Suite. In particular, the magnesite has been the focus of extensive prospecting, mapping, trenching, drilling, geochemistry and geophysics by several companies.

Zwicker and Strong (1986) reported significant gold values from mineralized boulders in Chiouk Brook. Subsequently, the Chiouk Brook gold showing was optioned from L. Murphy by U.S. Borax (Burton, 1987). The area was prospected, and geophysical, geochemical and drilling programs carried out. The same data were also reported in Mercer (1988a,b). Mineralized boulders located in the brook gave a high gold value of 0.6 oz./ton gold (approximately 18 g/t). They also reported that core assays indicated limited Au enrichment (maximum 2.2 g/t over 1 m) with local antimony and arsenic anomalies.

Falconbridge Limited (Vaskovic, 1987; Butler, 1988) examined the ultramafic suite adjacent to the main highway and reported anomalous Ni and minor Au values.

Mercer (1988a,b) reported that the Lizard Pond prospect gave generally much higher values than the Chiouk Brook prospect with one 50-m-long intersection averaging over 0.5 g/t with a high value of 2.6 g/t. The high gold values are commonly in fine-grained, silicified, sulphide-bearing magnesite and are accompanied by enhanced antimony and arsenic values.

The Spruce Brook Formation and the Great Bend Complex were subsequently staked by L. Murphy and optioned by B.P. Resources Canada Limited and BP Canada Limited (Graham, 1989a,b, 1990). Their extensive exploration work including drilling, trenching, geophysics, bedrock mapping and assaying. The reports are currently confidential.

Noranda Exploration Company Limited (Tallman, 1989, 1990, 1991) reported the discovery and geology of widespread antimony (stibnite) mineralization in the Cooper Brook area (Figure 1). Tallman (1991) reported that trenching and drilling indicated that stibnite and stibnite–quartz veins occur in coarse sandstone, pebble conglomerate and siltstone (Unit 29) at the Hunan deposit. A grade of 35.6 percent Sb over 1 m from a trench is the highest grade reported. Spectacular specimens of bladed antimony with the 20-cm-long blades forming rosettes have been obtained from the showing. The Xingchang showing is described as highly fractured graphitic shale and siltstone cut by stibnite and stibnite–quartz veins with an intersection of 33 percent Sb over 1 m in drill core.

A 15-m-long trench located 1.5 km west-northwest of Spud Pond was discovered to contain massive arsenopyrite veins in brecciated serpentinite (Figure 1). This trench was apparently dug by BP Resources Canada Limited in 1990. The occurrence is close to the faulted eastern contact of the serpentinite (Unit 1) with sandstone (Unit 11). A high-grade sample of arsenopyrite contains 10.1 percent arsenic and 1.1 g/t gold and 579 g/t antimony (this study). A channel sample gave 0.304 g/t Au, 3.88 percent As, and 228 g/t Sb. No other new significant metallic mineralization has been observed.

The building-stone potential of gabbro from the Mount Peyton Intrusive Suite has been studied for several years (Tomlin, 1982) with some recent positive impressions of the quality of the stone (Meyer, 1991). The gabbro in the map area does not appear to be of good quality because of closely spaced fractures, xenoliths and textural variations. The gabbro of Unit 16 is light coloured and contains widely spaced joints. Large piles of massive blocks are found near the woods roads. The pyroxenite (Unit 2) exposed along the road to Great Gull River consists of black, fresh massive rock with a complete absence of light-coloured minerals. Blasted blocks range in size up to 10 m^3. The extent of the fresh, massive rock is uncertain because of the thick till cover.

SUMMARY

The Eastern Pond (west half) map area contains the highly faulted Great Bend Complex that forms part of an ophiolite suite. The Complex is in faulted (probably overthrust) contact with the Early Ordovician Spruce Brook Formation. A 1-km-wide zone of steeply dipping, polydeformed mylonitic and amphibolite schist may represent the basal aureole of the suite. It also appears that possible Silurian sedimentary rocks are also imbricated with the ultramafic rocks.

The eastern part of the map area is underlain by turbiditic slate and sandstone of the Ordovician Davidsville Group. The lower parts of the Botwood Group consists of turbiditic sandstones and siltstone that are overlain by shallow marine to intertidal, sandstone and siltstone of the Silurian Botwood Group. A different succession of shallow marine to terrestrial Botwood Group sedimentary rocks underlies the northwestern part of the map area. The Mount Peyton Intrusive Suite is a composite, postkinematic intrusion of gabbro, tonalite, granodiorite and granite. Preliminary dating of pegmatitic gabbro indicates a Silurian age. Two other gabbroic units are possibly related to the suite. Three units of postkinematic, mainly coarse-grained, sedimentary rocks unconformably overlie the Ordovician chert, the Botwood Group, and a small part of the ultramafic suite.

The southern half of the ultramafic suite is extensively altered to magnesite and most significant gold mineralization occurs in silicified magnesite within this zone. Massive arsenopyrite mineralization occurs in brecciated serpentinite. The Chiouk Brook gold showing contains anomalous gold values in ultramafic rocks discovered in drill core. The surface outcrops of siltstone and diabase may represent a melange overlying imbricated ultramafic and sedimentary rocks. Antimony mineralization occurs in graphitic slate of the Davidsville Group and also the postkinematic sandstone and conglomerate at the Xingchang and Hunan deposits, respectively.

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PROBING THE CRUST WITH PLUTONS: REGIONAL ISOTOPIC GEOCHEMISTRY OF GRANITOID INTRUSIONS ACROSS INSULAR NEWFOUNDLAND

B.J. Fryer, A. Kerr[1], G.A. Jenner and F.J. Longstaffe[2]
Department of Earth Sciences
Memorial University of Newfoundland, St. John's, NF, A1B 3X5

ABSTRACT

This report presents an overview of Nd and O isotopic data from plutonic suites across the geographical and temporal spectrum of the Appalachian Wilson Cycle in Newfoundland, with emphasis on the Paleozoic suites. For the purposes of discussion, Newfoundland plutonic suites are divided into pre-orogenic (> 550 Ma) early orogenic (550 to 460 Ma) late orogenic (460 to 410 Ma) and post-orogenic (< 400 Ma) groupings.

Early orogenic suites, probably representing the products of arc magmatism within Iapetus, have variable but generally positive ϵ_{Nd} *(+1 to +6). Variations amongst these suites may be related to the original relative positions of different arc systems, with those closest to continental regions containing a higher proportion of older crust, or to the ages of the arcs and the material subducted beneath them. In strong contrast, ca. 460 Ma granites in south-central Newfoundland, with* ϵ_{Nd} *down to -8, were derived by melting of Gander Group supracrustals ± underlying basement, probably in response to tectonic accretion of the eastern Dunnage Zone.*

Late orogenic suites show strong isotopic variations that match their contrasts in petrology and elemental geochemistry. In the eastern part of the Central Mobile Belt, they range from S-type granites with strongly negative ϵ_{Nd} *(-5 to -8) and high* $\delta^{18}O$ *(>9) to metaluminous suites with* ϵ_{Nd} *around -3. The strongly negative granites were probably derived largely by melting of Gander Group supracrustal rocks. A regional decrease in* ϵ_{Nd} *from northeast to southwest further implicates the Gander Group, which is most abundant in the northeast, as partly responsible for these signatures. However, the persistent negative signatures on the south coast, where supracrustal rocks are rare, imply a Precambrian basement, which may also underly the northeastern Gander Zone at depth. In the northwestern Dunnage Zone, Silurian plutonic suites have largely juvenile signatures (+3 and above) that argue against the presence of Grenvillian basement rocks, and may indicate the presence of a juvenile (Paleozoic ?) lower crustal block, perhaps representing accreted arc systems. Much more work is required to confirm, characterize, and delineate this region.*

Post-orogenic suites also show very significant geographic isotopic variations. In the Avalon Zone, they have generally positive ϵ_{Nd} *values of +2 to +4 that argue against the presence of any ancient basement. In the Gander Zone, they show regional isotopic trends that resemble those of the late orogenic suites, and similarly suggest an input from the Gander Group supracrustal rocks. However, Devonian granites on the south coast still have negative* ϵ_{Nd}*, and contrast with those of the adjacent Avalon Zone. This suggests fundamental contrasts in the deep geology of these regions, and reinforces interpretation of the Dover–Hermitage Bay fault system as a major structural break. Thus, late Precambrian rocks on the south coast may not correlate directly with the 'type' Avalon Zone, although they may both be part of a single composite Precambrian terrane representing the Gondwanan side of the Orogen.*

The isotopic data are by no means complete, and further work is required to both confirm and expand some of the more tentative inferences. However, despite these and other problems, the regional isotopic data provide valuable information with which to delineate and characterize major components of the Appalachian Orogen in Newfoundland. The challenge for the final stages of the LTIHOPROBE EAST granite project is to complete the isotopic database and successfully integrate it with other relevant geophysical and geological data.

Lithoprobe contribution # 287

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[1] Newfoundland Mapping Section, Geological Survey Branch, Department of Mines and Energy

[2] Department of Geology, University of Western Ontario, London, ON, N6A 5B7

INTRODUCTION

LITHOPROBE EAST is a major multidisciplinary project that aims to delineate and understand crustal and subcrustal structure within the Appalachian Orogenic Belt. Seismic reflection profiling to date has included marine profiles to the northeast and southwest of Newfoundland (Keen *et al.*, 1986; Marillier *et al.*, 1989), and, more recently, onshore Vibroseis profiles (Quinlan *et al.*, 1991). Supporting geoscience projects, in which surface-based studies seek to augment the seismic database, include a regional isotopic and geochemical study of granitoid intrusive rocks across Newfoundland (Fryer *et al.*, 1989). This is coordinated with a Geological Survey Branch (GSB) project intended to compile, augment and interpret geochemical data from these rocks (Kerr *et al.*, 1991). As granitoid plutons are at least partly derived from the deep continental crust, they provide 'windows' through which the features of lower regions imaged by seismic data may be glimpsed. Also, isotopic data from plutonic suites distributed across the spatial and temporal spectrum of an orogenic belt provide insight into the sources of granitoid magmas and their relationship to orogenic evolution.

Although a large number of isotopic analyses are now complete, these data have previously been reported only in abstract or outline form (e.g., Fryer *et al.*, 1989; Kerr *et al.*, 1990). This report is a regional overview of results and their geological and tectonic implications. It concentrates mostly upon neodymium and oxygen isotope data.

With respect to this account, two points are stressed. First, the data are not yet complete; new information may neccessitate revision of some more tentative ideas presented here. Second, data are derived from field work and sampling by several researchers whose names do not appear in the list of authors. The purpose of this contribution, as stated earlier, is to provide an overview, and to highlight some specific problems for further attention. It is anticipated that separate papers will eventually address more localized and focussed problems.

REGIONAL GEOLOGICAL AND GEOPHYSICAL FRAMEWORK

UPPER CRUSTAL SUBDIVISIONS DEFINED BY SURFACE REGIONAL GEOLOGICAL PATTERNS

Newfoundland is one of the best type-areas for Appalachian–Caledonian orogenic evolution, as it displays a relatively complete cross-section of the belt, and preserves the largest remnants of the Iapetus Ocean and/or its associated basins. The current 'plan view' of the Newfoundland Appalachians (Figure 1) comprises four main tectonostratigraphic zones (Williams, 1979; Williams *et al.*, 1988).

From west to east, these are the *Humber Zone* (the ancient North American continental margin), the *Dunnage Zone* (various remnants of the Iapetus realm, mostly comprising island arcs and back-arc basins), the *Gander Zone* (a metamorphic belt of sedimentary protolith possibly representing an opposing, Gondwanan, continental margin) and the *Avalon Zone* (a Late Proterozoic orogenic province with a thin Paleozoic cover sequence; probably a fragment of Gondwanaland developed during the Pan-African cycle). The Dunnage Zone is subdivided into two major segments, termed the *Notre Dame Subzone* and *Exploits Subzone*, which have distinct lithological, faunal and evolutionary characteristics (Williams *et al.*, 1988). The zonal scheme encounters problems on the south coast which, although traditionally placed within the Gander Zone, lacks the metasedimentary rocks that are characteristic of this division. In Figure 1, it has been retained as part of the Gander Zone, mainly because its plutonic evolution appears similar to areas farther to the northeast, and they are most conveniently discussed together.

Colman-Sadd and Swinden (1984) proposed that the eastern Dunnage Zone was allochthonous, and that metasedimentary rocks of the Gander Zone formed tectonic 'windows' within it. This concept is also supported by gravity data (Karlstrom, 1983). Marine seismic profiles (Keen *et al.*, 1986; Marillier *et al.*, 1989) also suggest that thick continental crust everywhere underlies the Dunnage Zone.

From the perspective of Paleozoic plutonism, the most important portion of the orogen is the *Central Mobile Belt* (i.e., the Dunnage and Gander zones combined), which is the locus of voluminous magmatism (Figures 1 and 2). Paleozoic plutons (of post-Silurian age) also occur in the Avalon Zone, but there are virtually none in the Humber Zone, except at its easternmost edge. Proterozoic granitoid rocks, however, are important components of both the Humber and Avalon zones, where they are manifestations of the earlier Grenvillian and Pan-African cycles, respectively (Williams *et al.*, 1989).

LOWER CRUSTAL SUBDIVISIONS DEFINED BY DEEP SEISMIC REFLECTION PROFILES

The results of marine seismic reflection profiles are discussed by Keen *et al.* (1986) and Marillier *et al.* (1989); further discussion and interpretation are presented by Stockmal *et al.* (1990). These results are interpreted in terms of three lower crustal blocks (Figure 1), defined by contrasts in the depth of the Moho, and the attitudes, positions and abundances of mid- to lower-crustal reflectors. The boundaries of these blocks are interpreted to partially coincide with surface geological boundaries, based on along-strike extrapolation using regional geophysical data. Figure 1 shows the distribution of blocks according to Marillier *et al.* (1989); however, the regional isotopic data presented in this paper are not fully supportive of this pattern.

The *Grenville Block* was interpreted to underlie the Humber Zone and at least part of the Dunnage Zone, and the *Avalon Block* was interpreted to coincide with the surface expression of the Avalon Zone. The intervening region was termed the *Central Block;* this underlies the Gander Zone

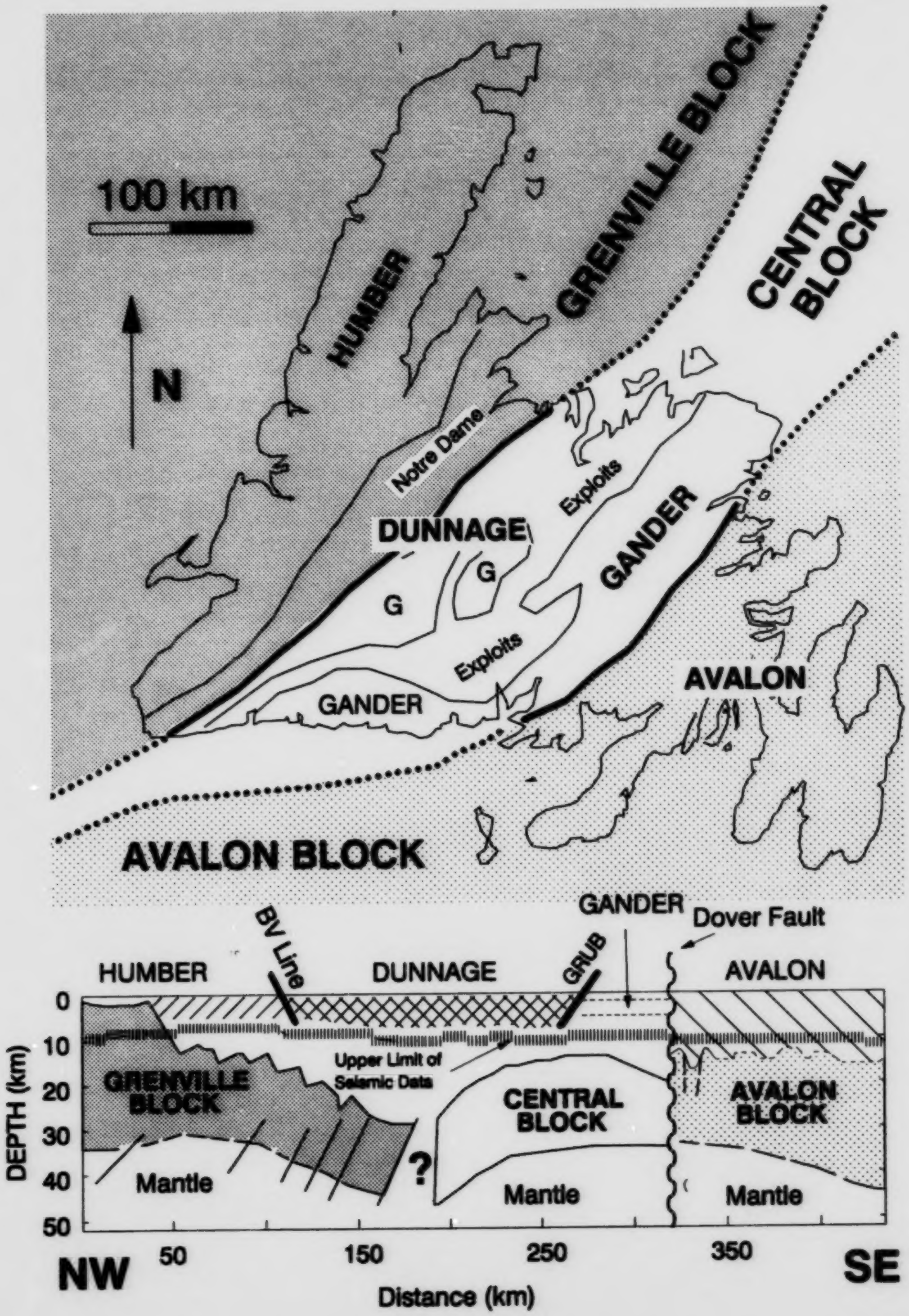
100 km
N
HUMBER
GRENVILLE BLOCK
CENTRAL BLOCK
Notre Dame
DUNNAGE
Exploits
G
G
GANDER
Exploits
GANDER
AVALON
AVALON BLOCK
BV Line
GANDER
Dover Fault
HUMBER
DUNNAGE
GRUB
AVALON
Upper Limit of Seismic Data
GRENVILLE BLOCK
CENTRAL BLOCK
AVALON BLOCK
?
Mantle
Mantle
Mantle
DEPTH (km)
0
10
20
30
40
50
NW
50
150
250
350
SE
Distance (km)

Figure 1. *Major subdivisions (tectonic zones) of the Newfoundland Appalachians (after Williams, 1979; Williams* et al., *1988) in relation to lower crustal blocks postulated from seismic transects (Keen* et al., *1986; Marillier* et al., *1989). The schematic cross-section is based on these, and other sources.*

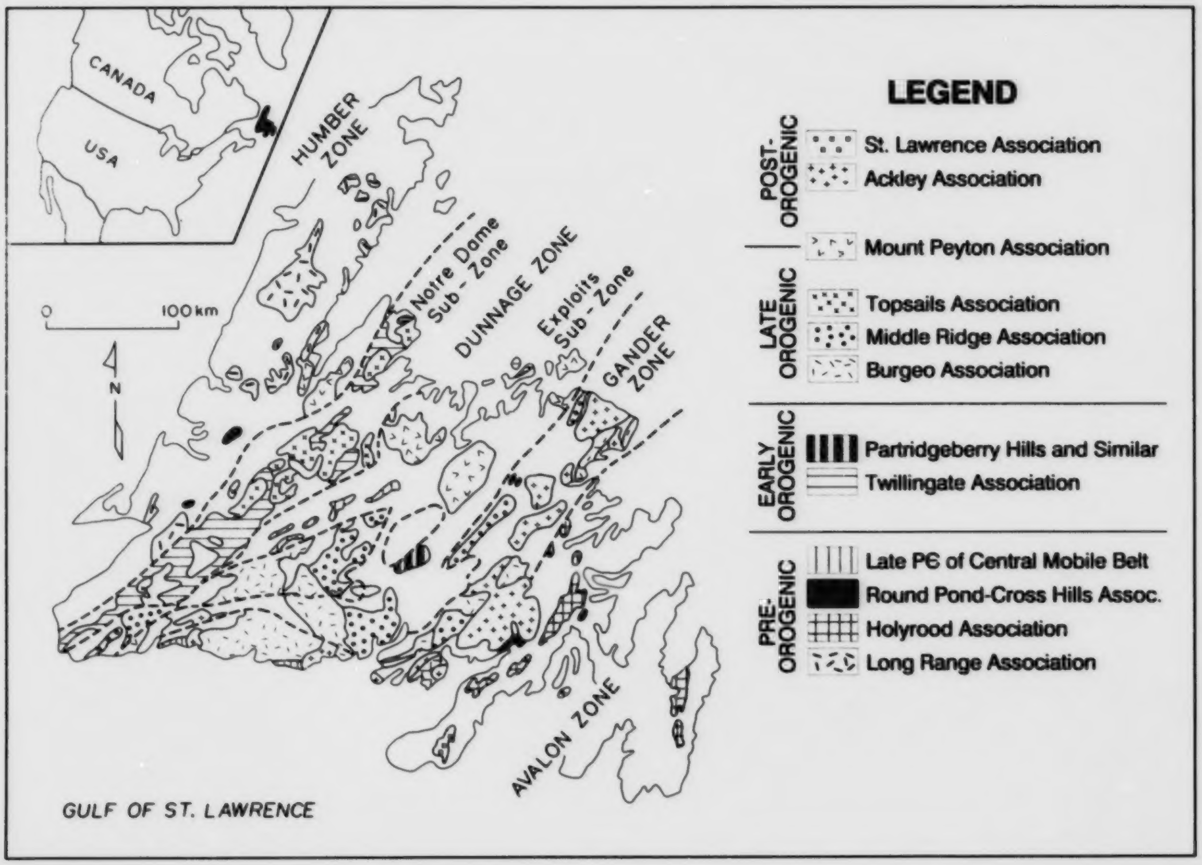
CANADA
USA
HUMBER ZONE
Notre Dame Sub-Zone
DUNNAGE ZONE
Exploits Sub-Zone
GANDER ZONE
AVALON ZONE
0
100km
N
GULF OF ST. LAWRENCE
LEGEND
POST-OROGENIC
St. Lawrence Association
Ackley Association
Mount Peyton Association
LATE OROGENIC
Topsails Association
Middle Ridge Association
Burgeo Association
EARLY OROGENIC
Partridgeberry Hills and Similar
Twillingate Association
PRE-OROGENIC
Late PЄ of Central Mobile Belt
Round Pond-Cross Hills Assoc.
Holyrood Association
Long Range Association

Figure 2. *Distribution and classification of plutonic suites in the Newfoundland Appalachians, after Williams* et al. *(1989) with modifications to account for post-1989 research (Dunning and O'Brien, 1989; Evans* et al., *1990; Colman-Sadd* et al., in press*).*

and much of the Dunnage Zone. The presence of thick continental crust beneath central Newfoundland implies that the oceanic terranes of the Dunnage Zone are allochthonous, with a maximum vertical extent of around 12 km (Keen *et al.*, 1986). There has been substantial discussion of the seismic interpretation, especially since the acquistion of on-land profiles (Quinlan *et al.*, 1991). However, the tripartite division outlined above remanins generally applicable in the context of this paper.

Of the three lower crustal blocks imaged by seismic reflection, only the Grenville can be studied directly. There are no exposures of 'basement' within the Avalon Zone (at least in Newfoundland), and little unequivocal sialic basement within the Gander or eastern Dunnage zones. Indirect methods are thus required to constrain their characteristics beyond the physical and structural parameters provided by seismic data. Also, any inferences from offshore profiles depend heavily upon lateral correlations of 100 km or more, and a more precise analysis of correspondance between surface and deep-crustal geology requires a method of mapping their distribution *in situ*. The onshore seismic profiles (Quinlan *et al.*, 1991) provide a partial solution to this problem, but still provide only two-dimensional vertical profiles of the crust in most areas.

GRANITOID INTRUSIONS AS WINDOWS TO THE LOWER CONTINENTAL CRUST

Plutonic rocks, particularly those of granitoid composition, provide a useful method of mapping the subsurface distribution of crustal blocks. Although there is debate about the ultimate sources of granitoid magmas, most are derived, at least in part, from rocks of the lower continental crust. Consequently, they preserve both isotopic and elemental geochemical 'images' of these lower crustal sources, and can define and delimit basement provinces and their boundaries (e.g., DePaolo, 1981; Farmer and DePaolo, 1983; Halliday, 1984; Kerr and Fryer, 1990). In addition to 'mapping' such blocks, the geochemical signatures of granitoid rocks can provide information about their age and affinities, and, perhaps most importantly, may be able to identify and delineate regions that cannot be distinguished with confidence on the basis of seismic character alone.

The interpretation of such patterns is, however, unlikely to be straightforward. Some have argued that granites provide direct images of such sources because they are derived wholly by crustal anatexis (e.g., Chappell and Stephens, 1988; Chappell *et al.*, 1988), whereas others have suggested complex blending of crustal and subcrustal (i.e., mantle-derived) components (e.g., DePaolo, 1981; Harmon *et al.*, 1984; Hildreth and Moorbath, 1988). A second problem is that granitoid magmas may be derived from contrasting crustal levels; e.g., Al-rich 'S-type' granites, commonly viewed as anatectic derivatives of sedimentary rocks (e.g., Chappell and White, 1974; White and Chappell, 1990), may not have any direct relationship to their underlying basement rocks. A third problem is that there is no requirement for interfaces between crustal blocks to be vertical, and contrasting provinces could be juxtaposed in a vertical sense by, for example, thrusting. However, in spite of these and other problems, granitoid plutonic rocks provide 'windows' through which the features of the lower crust can be visualized and mapped, and their study is a crucial component of LITHOPROBE EAST.

GRANITOID PLUTONIC ASSOCIATIONS IN NEWFOUNDLAND: AN OVERVIEW

About 35 percent of Newfoundland is underlain by rocks belonging to the granitoid clan, ranging in composition from diorite to alkali-feldspar granite, with associated mafic intrusive rocks (Figure 2). A major contribution to their study was made by D.F. Strong in conjunction with survey mapping projects (e.g., Strong *et al.*, 1974; Strong and Dickson, 1978; Payne and Strong, 1979; Strong, 1980). During the 1980's, systematic mapping has added greatly to our petrological and geochemical database (e.g., Dickson, 1983, 1990; Whalen *et al.*, 1987), and U−Pb zircon geochronology has provided a more tightly constrained temporal framework, (e.g., Krogh *et al.*, 1985; Dunning and O'Brien, 1989; Dunning *et al.*, 1987, 1990, 1991). The following overview is, as far as is possible, a 'state-of-the-art' summation, but it should be noted that many plutons are still not dated with precision; work in progress by G.R. Dunning and R.D. Tucker, in conjunction with GSB mapping projects, may result in modifications of temporal groupings.

In this paper, we adopt the generalized descriptive classification outlined by Williams *et al.* (1989), based on earlier schemes presented by Strong (1980) and Hayes *et al.* (1987). Several 'types' or associations of plutonic rocks are defined on the basis of age, petrological features and relationships to tectonic zonations, and named after prominent examples (Figure 2). The groupings are consistent with overall geochemical patterns (Dickson *et al.*, 1990; Kerr *et al.*, 1990), although geochemistry alone cannot provide absolute discrimination between associations. There are some newly recognized associations that do not readily fit into the scheme, such as Ordovician S-type granites in the Gander Zone (Colman-Sadd *et al.*, *in press*) and late Precambrian intrusions within the Dunnage Zone (Evans *et al.*, 1990), and it is viewed as an evolving, rather than a rigid, classification scheme.

In this paper, the descriptive classification of Williams *et al.* (1989) is overlain by a temporal framework, in which Newfoundland granitoid suites are divided into *pre-orogenic, early orogenic, late orogenic,* and *post-orogenic* categories. There is a general, but not exact, correspondance between these and the descriptive associations illustrated in Figure 2.

Two points need to be stressed with regard to this framework. First, these terms are used with reference to the *Appalachian Orogenic Cycle;* thus, pre-orogenic intrusions may be syn-, post- or anorogenic with respect to earlier events such as the Grenvillian or Pan-African orogenic cycles. Second, the use of the term 'orogenic' does not imply a single, constrained event (as in 'Taconic Orogeny'), but rather the protracted and/or episodic tectonism that accompanied and followed the destruction of Iapetus. This usage is akin to the definition used by geologists working in Precambrian shield areas; for specific deformational markers, the term event is used here. In terms of stratigraphic terminology, these groupings are as follows:

Pre-orogenic = Proterozoic or Eocambrian (> 550 Ma),
Early Orogenic = Cambrian and Earlier Ordovician (550-460 Ma),
Late Orogenic = Later Ordovician and Silurian (460-410 Ma),
Post-orogenic = Later Silurian and younger (410-360 Ma)

These categories are not discrete 'pigeonholes', and their context is that of the Newfoundland Appalachians alone, but they form a useful framework for discussion of the isotopic data. The first group has only a tenuous connection to the Appalachian Wilson Cycle, but is of importance in that these materials may have contributed to younger magmas. These temporal groupings are summarized below; their individual components and features are discussed in subsequent sections of the report, in conjunction with the isotopic data.

Pre-orogenic Granitoid Associations

This grouping includes the Long Range, Holyrood, Round Pond and Cross Hills associations of Williams *et al.* (1989). The Long Range (≈1000 Ma) and Holyrood (≈620 Ma) associations are related to the Grenvillian and Pan-African orogenic cycles respectively, and were emplaced mostiy after final deformational events in each. They are dominated by calc-alkaline to alkali-calcic, hornblende-bearing granodiorites to granites. The Round Pond (≈610 Ma) and Cross Hills (≈560 Ma) associations comprise small, variably alkaline plutons that intrude the Humber and Avalon zones respectively (Figure 2). Williams *et al.* (1989) suggested that the Cross Hills and Round Pond associations may reflect rifting to form Iapetus.

Granitoid suites of late Precambrian age also occur on the south coast and in central Newfoundland. The U−Pb ages of those on the south coast are broadly analogous to those of the 'type' Avalon Zone, and lead to suggestions that tectonostratigraphic zones should be revised in this region

(Dunning and O'Brien, 1989; O'Brien *et al.*, 1991). Deformed granites close to the Exploits–Notre Dame subzone boundary have also yielded U–Pb ages of ca. 560 Ma (Evans *et al.*, 1990). These may represent exhumed Precambrian basement that occurs at depth beneath this region. Late Precambrian intrusions in both these regions are grouped together in Figure 2.

Early Orogenic Granitoid Associations

The most widespread of these is the Twillingate Association of Williams *et al.* (1989), including a variety of generally sodic (tonalitic, trondhjemitic and granodioritic) intrusive complexes ranging in age from 510 Ma to about 460 Ma, most of which were affected by the Taconic event. Some are demonstrably allochthonous (e.g., the Little Port Complex in the Humber Arm Allochthon), and most are probably allochthonous on a crustal scale. The most extensive tonalitic granitoid terrane is in southwestern Newfoundland (Figure 2), an area once referred to poetically as a 'sea of tonalite'. Some Ordovician ages have been obtained (e.g., Dunning *et al.*, 1989), but the plutonic suites of this area remain poorly known.

These early orogenic tonalite–granodiorite suites have been widely interpreted to result from subduction-related magmatism broadly associated with island- and/or continental-arc settings (e.g., Williams and Payne, 1975; Dunning *et al.*, 1987, 1989), and they record the destruction of at least parts of Iapetus or related basins. Recent geochronological data from arc-type volcanic sequences (Dunning *et al.*, 1991) indicates that subduction commenced prior to about 513 Ma. The probable allochthonous character of these early orogenic granitoid suites implies that their geochemical and isotopic signatures may be unrelated to the lower crustal block(s) that *presently* underlie(s) the Dunnage Zone. Variations in age and petrology amongst these suites suggest that they almost certainly represent the products of more than one magmatic belt or episode, and there is every reason to expect isotopic variations linked to geography or age.

Until recently, it was considered that Ordovician plutonic suites were absent from eastern Newfoundland. However, recent U–Pb geochronology has shown that at least three intrusions within the Gander Zone crystallized between 470 and 460 Ma (Colman-Sadd *et al.*, *in press*). These are leucocratic, locally muscovite-bearing granites that are distinct from the arc-type suites noted above. These are grouped separately in Figure 2; not all examples can be shown at the scale of the map.

Late Orogenic Granitoid Associations

This broad group comprises granitoid suites emplaced during the later Ordovician and Silurian, between about 460 and 410 Ma. Silurian magmatism and related deformation (Salinic Event) is of great importance in Newfoundland (e.g., Dunning *et al.*, 1990). The most important feature of these late orogenic suites is their marked compositional variation across the island, from west to east; the two regions are described separately below.

In the Notre Dame Subzone of the Dunnage Zone, most post-Ordovician intrusions were grouped by Williams *et al.* (1989) into the *Topsails Association* (Figure 2). The type example of this grouping is a large, metaluminous to peralkaline complex of ca. 430 to 415 Ma (Whalen *et al.*, 1987; Whalen, 1989); similar suites occur in the Springdale area and on the Baie Verte Peninsula (Coyle and Strong, 1987; Hibbard, 1983). These intrude volcanic suites of the Notre Dame Subzone, and the foliated early orogenic plutonic rocks (see above). All are variably bimodal, and most include minor mafic plutonic rocks. In a local context, these are dominantly posttectonic, being younger than the Taconic event. Williams *et al.* (1989) also included some intrusions of the easternmost Humber Zone in this group; however, Kerr *et al.* (1990) argued that geochemical data did not support this correlation. With this modification, Topsails Association plutons are entirely restricted to the Notre Dame Subzone, and probably reflect a distinct tectonic or thermal history for this region (Kerr *et al.*, 1990).

The other main locus of Silurian magmatism is in the eastern Central Mobile Belt (shown mostly as Gander Zone in Figures 1 and 2); a few examples also cut the structurally overlying Exploits Subzone. Williams *et al.* (1989) distinguished two groupings in the Gander Zone, which they termed the *Burgeo* and *Middle Ridge associations* (Figure 2). These are radically different from the Silurian suites of the western Dunnage Zone; they are dominated by metaluminous to variably peraluminous biotite- and two-mica granite suites, which intrude clastic metasedimentary rocks of the Gander Group and similar or correlative sequences. Their relationship to local deformational events is variable, and many appear to be broadly syn-tectonic, implying a major Silurian Orogenic pulse (e.g., Dunning *et al.*, 1990).

The *Burgeo association* is dominated by variably foliated, K-feldspar megacrystic, biotite granites, which are concentrated along the eastern fringe of the Gander Zone in northeastern Newfoundland, and on the south coast, in areas of relatively high metamorphic grade. The type example is a complex, polyphase batholith on the south coast; similar rock types can be traced as far as Cape Freels, west of Bonavista Bay. Subordinate biotite–muscovite granites also occur in this association, and, based on field relationships and U–Pb dates, appear to be slightly younger than the megacrystic granites, and are possibly closer in affinity to the Middle Ridge association (see below). Tonalitic and dioritic variants are also reported from the type example (Dickson *et al.*, 1989). The *Middle Ridge association* is mostly localized in interior, lower-grade areas (including parts of the Exploits Subzone) and comprises leucocratic two-mica granites having a more restricted compositional spectrum. The type example has been dated at 410 Ma (U–Pb monazite; Tucker, 1990), but reliable geochronological data are sparse elsewhere.

In compositional terms, the Burgeo and Middle Ridge associations form a continuum (Dickson *et al.*, 1990; Kerr *et al.*, 1990) from metaluminous to weakly peraluminous (Burgeo) to strongly peraluminous (Middle Ridge) suites. The Middle Ridge association exhibits many of the characteristic geochemical features of S-type granites, suggested to be derived by anatexis of metasedimentary protoliths (Chappell and White, 1974; White and Chappell, 1990).

Kerr *et al.* (1990) suggested that the features of late-orogenic suites in eastern Newfoundland might reflect *depth zonation*, with the Burgeo Association representing a deep-level magmatic environment, whereas the Middle Ridge Association was derived largely from metasedimentary materials higher in the crust, but below their current level of exposure. From the perspective of isotopic studies aimed at lower crustal blocks, inferences from the latter should perhaps be treated with caution, as they may sample a fairly restricted source region of the crust, which itself may reflect a variety of sources that have no direct connection with the basement presently beneath the area.

Post-orogenic Granitoid Associations

Post-orogenic magmatic associations, of Devonian and younger age, fall into two main groups.

The most easily defined is the *Ackley association* of Williams *et al.* (1989), comprising a major belt of posttectonic plutons of 400 to 380 Ma age stretching from the south coast to Bonavista Bay (Figure 2). The Ackley association occurs within the Gander Zone (as indicated on Figure 2) and the westernmost Avalon Zone, suggesting that these components of the orogen were juxtaposed prior to their emplacement. These plutons are dominated by siliceous, potassic biotite–monzogranite and granites, with both K-feldspar megacrystic (notably in the northeast) and equigranular variants (Dickson, 1983; Dickson *et al.*, 1989). Associated Mo–Sn–W mineralization, and ring-complex to lobate geometry, suggest relatively high emplacement levels. Williams *et al.* (1989) included the peralkaline St.Lawrence Granite (Figure 2) as part of this association. This pluton has been dated as Carboniferous (326 Ma; Bell *et al.*, 1977), but this age is likely disturbed; U–Pb zircon dating is currently underway. As the St. Lawrence Granite is geochemically unique amongst Paleozoic plutons in eastern Newfoundland, it is retained in a separate grouping, for the present, and is not discussed further.

Williams *et al.* (1989) defined the *Mount Peyton Association* to include the bimodal Late Silurian to Devonian plutons of the Dunnage Zone, concentrated mostly in the Exploits Subzone. The type example (Strong, 1978) includes extensive gabbroic to dioritic rocks, and a younger metaluminous biotite-granite; similar diorite–granite associations typify the Fogo Island and Hodges Hill complexes. The Devonian plutonic suites of the easternmost Humber Zone (formerly grouped with the Topsails) are tentatively included as part of this association. These bimodal suites are generally rather poorly known, and this subdivision has commonly become a resting place for any plutonic rocks of uncertain affinity.

The K–Ar and Ar–Ar dates suggest that the Mount Peyton Association may straddle the rather arbitrary late orogenic and post-orogenic temporal groupings used here. Mafic components typically appear of Silurian (> 420 Ma) age, whereas the granites (*ss*) yield Devonian ages; however, the data are not comprehensive at present, and U–Pb data are required to substantiate this pattern. The spatial association of mafic and felsic rocks suggests a genetic link, perhaps indirect, but it is not yet clear if they have similar ages of emplacement. We have elected to represent most of these as Devonian in our isotopic variation maps (Figures 4 to 7), but this may be subject to revisions.

ND AND O ISOTOPIC DATA: AN OVERVIEW OF THEIR USAGE

In this section, we present a brief explanation of the basic principles of Sm–Nd and O isotope geochemistry (after DePaolo, 1988; Hoefs, 1981). This is intended to provide background information relevant to discussions of data that follow, and is by no means comprehensive.

SM–ND ISOTOPIC SYSTEM

Basic Principles

This radiogenic decay system is analogous to the more familiar Rb–Sr system, but different in important respects. The LREE parent (^{147}Sm) and daughter (^{143}Nd) elements have closely similar ionic radii and valence properties, unlike Rb and Sr. Thus, Sm and Nd are not easily separated by crustal processes such as anatexis or fractional crystallization, and the range of Sm/Nd ratios in crustal rocks is very small (0.1 to 0.3) compared to Rb/Sr ratios (< 0.01 to > 100).

The Nd isotopic variations in nature arise mostly in the partial melting of the mantle to generate crustal rocks, when the mantle retains Sm preferentially over Nd. Over geological time scales, the upper mantle has been depleted in Nd, and this 'depleted mantle reservoir' has evolved to have higher ^{143}Nd/^{144}Nd ratios. Conversely, crustal rocks are relatively depleted in Sm, and have evolved to have significantly lower ^{143}Nd/^{144}Nd ratios. Thus, if the initial ^{143}Nd/^{144}Nd ratio of a rock is known, it will reveal whether that rock formed directly from a mantle source, or includes some older pre-existing crustal source (Figure 3). The existence of the depleted mantle reservoir is confirmed by isotopic studies of mafic volcanic rocks through geological time (DePaolo, 1988).

Notation and Representation

The actual variation of ^{143}Nd/^{144}Nd in nature is very small, and a simple scaling technique is employed to give a parameter termed ϵ_{Nd}:

$$\epsilon_{Nd} = [(^{143}\text{Nd}/^{144}\text{Nd} \;/\; ^{143}\text{Nd}/^{144}\text{Nd CHUR}) - 1] \times 10^4$$

Here, CHUR ('chondritic uniform reservoir') is a chondritic value, assumed to be representative of the bulk

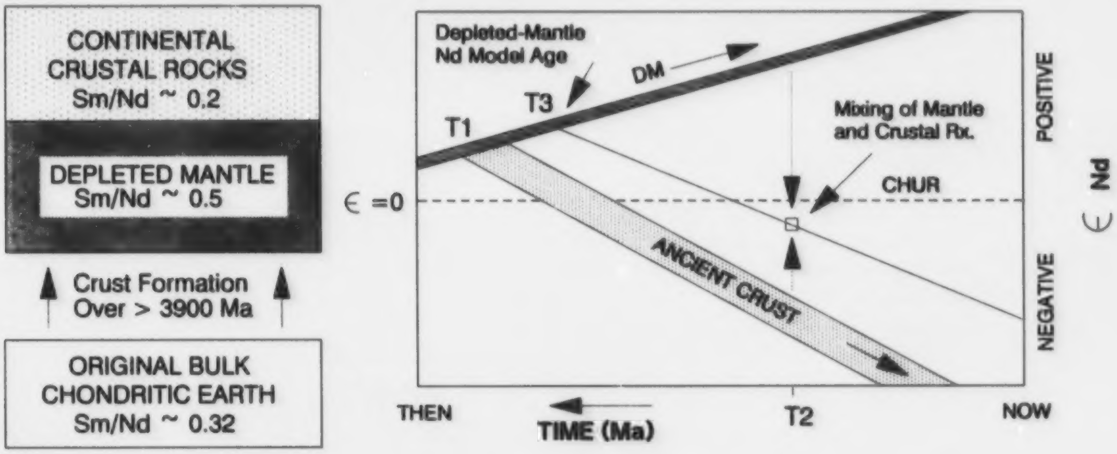
CONTINENTAL CRUSTAL ROCKS Sm/Nd ~ 0.2
DEPLETED MANTLE Sm/Nd ~ 0.5
Crust Formation Over > 3900 Ma
ORIGINAL BULK CHONDRITIC EARTH Sm/Nd ~ 0.32
Depleted-Mantle Nd Model Age
DM
T3
T1
Mixing of Mantle and Crustal Rx.
CHUR
∈ =0
ANCIENT CRUST
POSITIVE
NEGATIVE
∈ Nd
THEN
TIME (Ma)
T2
NOW

Figure 3. *Schematic representation of Nd isotopic reservoirs, and their evolution in time, and the interpretation of Nd isotopic data. Partly after Arndt and Goldstein (1987); see text for explanation and discussion.*

Earth at formation. The ϵ_{Nd} value of any material changes with time as Sm continues to decay; thus, it is conventionally calculated for the geological age of the sample, i.e., both the Nd isotopic compositions of the sample and CHUR are corrected to 'initial' values. The ϵ_{Nd} scale is easy to use, as it generally varies between about -20 and +10; positive values are enriched in ^{143}Nd (usually indicative of mantle origin) and negative values are depleted in ^{143}Nd (usually indicative of an older crustal component).

Interpretation of Data

The ϵ notation enables simple graphical representations of Nd isotopic data and isotopic evolution, as depicted in Figure 3 (after Arndt and Goldstein, 1987). Here, the horizontal line represents a theoretical chondritic composition, which always has ϵ_{Nd}=0. The depleted mantle (Sm/Nd > CHUR) evolves with a positive slope, tending toward higher ϵ_{Nd} with time. Ancient crustal material (Sm/Nd < CHUR), derived from the mantle at time T1, evolves toward negative ϵ_{Nd}. If a hypothetical rock, known to have crystallized at time T2, has a slightly negative ϵ_{Nd} value at that time, it was not derived entirely from the mantle, and must include some older crustal component.

However, there are two ways of interpreting this result. The simplest interpretation is that the rock formed via anatectic 'reprocessing' of pre-existing crust. The time at which this precursor formed is found by extrapolating the growth line of the sample backward in time until it meets the depleted mantle line. This gives the *depleted-mantle model* age, or T_{dm} (T3). This represents the maximum age of the precursor material. An alternative interpretation is that the rock formed at time T1 as a homogenized mixture of a mantle-derived magma, and older crustal rocks. This is a very likely scenario, as many models for granitoid magma genesis call upon such processes. In this case, the crustal component is significantly older than T3, and the 'model age' of the rock has no real geological meaning. Resolution of these alternatives depends on other types of data, e.g., trace-element geochemistry or analysis of zircon populations. However, the most important point, is that, in either case, the Nd isotopic signature of the granite is a sensitive indicator of the presence of older crustal rocks. Also, in either case, the model age provides a maximum for the geological age of the sample.

It should be pointed out that minor fractionation of Sm from Nd by anatexis or other processes can distort evolutionary patterns from the straight-line models assumed in Figure 3. However, such effects are very small compared to other systems (notably Rb–Sr) and can be quantified using REE patterns. The S-type leucogranites, which may be derived from garnet-bearing metasedimentary sources, are the most prone to such disturbance. The miniscule variation of Sm/Nd ratios in nature indicates that this isotopic system is, in most cases, well-behaved.

OXYGEN ISOTOPES

Oxygen isotopes (^{18}O and ^{16}O) are stable, but, as they are light, they are subject to mass fractionation. Variations are expressed as the parameter δ^{18}O, calculated as:

$$\delta^{18}O = [(^{18}O/^{16}O\ /\ ^{18}O/^{16}O\ SMOW)-1] \times 10^3$$

SMOW is a standard value, referring to 'standard mean oceanic water'. This notation is analogous to the ϵ notation for Nd, except that δ^{18}O does not change with time, and the scaling factor is smaller. Variations in oxygen isotope compositions result from a wide range of processes, including evaporation, condensation, crystallization and diagenesis. Most sedimentary rocks, which have passed through the hydrosphere, have high ('heavy') δ^{18}O of > 10. Typically, granites have δ^{18}O from + 5 to + 12. Values of around + 6 resemble those of mantle-derived rock types, but values above + 9 resemble those of sedimentary rocks. Thus, δ^{18}O values are potentially useful tracers of sedimentary input into granitoid magmas.

ISOTOPIC DATA USED IN THIS PAPER

Nd and oxygen isotopic data were acquired over a three-year period, commencing in 1988. No numerical data listings are presented here.

Elemental concentrations of Nd and Sm are measured by inductively-coupled plasma mass-spectrometry (ICP-MS), using the methods of Jenner *et al.* (1990); overall uncertainties are estimated at ± 5 percent or better (2 sigma). The ^{147}Sm/^{144}Nd ratios used in this study are calculated from the concentration data above, and their overall uncertainties are estimated at ± 3 percent or better (2 sigma). The ^{143}Nd/^{144}Nd ratios are measured by thermal ionization mass spectrometry, after chemical separation of Nd from Sm and other REE by ion-exchange chemistry. Their overall uncertainty (2 sigma) is ± 0.007 percent or better. Calculations of ϵ_{Nd} at the time of formation (or at any other time) use the standard CHUR values; present-day ^{143}Nd/^{144}Nd = 0.512638 and ^{143}Nd/^{144}Nd = 0.196593. Depleted-mantle model ages are calculated according to the models of DePaolo (1988). The overall uncertainty in the calculated ϵ_{Nd} value for most samples is estimated at ± 0.5 ϵ units or better; in data representation, we show ϵ_{Nd} values to the nearest whole number. It should be noted that ϵ_{Nd} is sensitive mostly to ^{143}Nd/^{144}Nd ratios, and less sensitive to inaccuracies in the ^{147}Sm/^{144}Nd ratio. Also, the geological age of a sample is required to calculate ϵ_{Nd}, but the final value is insensitive to age variations of around ± 50 Ma. A 50 Ma age uncertainty translates to around ± 0.5 ϵ units for most samples; the younger the age, the lower the ϵ_{Nd} value.

Oxygen isotope compositions were measured by gas source mass-spectrometry at the University of Western Ontario. The overall uncertainty in the final δ^{18}O values calculated for most samples is estimated at ± 0.2 δ^{18}O units.

REGIONAL ND AND OXYGEN ISOTOPIC PATTERNS

In this section of the paper, the regional isotopic variations amongst specific groups of plutonic suites are depicted as maps, in which calculated ϵ_{Nd} or δ^{18}O values are shown. Some of these values represent single determinations; others are averages of more than one sample from the same location, or from closely spaced locations. For most plutons, 2 or more samples have been analyzed; where there are large variations between samples, the extreme values have been shown. As the symbols used are about 25 km in diameter, at the map scale, their positioning is approximate only, and we have not attempted to show the outlines of the plutonic suites that they represent.

PRE-OROGENIC PLUTONIC SUITES AND OTHER POTENTIAL SOURCE MATERIALS

Presently available isotopic data come mostly from suites that are related to the Appalachian cycle, and analyses of the plutonic rocks produced in the Grenvillian and Pan-African events are sparse. However, these rocks are important, as they are the only accessible representatives of potential basement blocks, and possible sources for younger Paleozoic plutonic suites. Similarly, the isotopic features of sedimentary sequences within the Central Mobile Belt (e.g., the Gander Group) are of interest, as these rocks may have been melted below the present erosion level to produce some of the granites that now intrude them.

Pre-orogenic Plutonic Suites of Western Newfoundland

The Nd isotopic compositions for these are represented at 430 Ma, rather than at their time of formation at around 1050 Ma (Figure 4a). This illustrates their features at a time when they may have been remelted to contribute to the younger Paleozoic magmas; in the context of this paper, this is probably more relevant than their original features. Early to Middle Proterozoic T_{DM} model ages of 1900 to 1500 Ma for most of these rocks are consistent with evidence that much of the Grenville Province in adjacent Labrador was initially formed during an earlier (Labradorian) orogenic cycle (e.g., Gower, 1991).

The Lake Michel Intrusive Suite, Potato Hill pluton and Long Range Inlier gneisses all have significantly negative ϵ_{Nd} values at 430 Ma, ranging from -5 to -12 (Figure 4a). Thus, any Paleozoic granite magma derived entirely by melting of materials similar to the Long Range Inlier (i.e., the Grenville Crustal Block) should have a strongly negative signature that reflects the antiquity of this source.

Oxygen isotopic compositions (Figure 4b) vary from δ^{18}O of +7 to +9 for most Grenvillian plutonic rocks to +11 in the more leucocratic rocks with strongly negative ϵ_{Nd}; Long Range Inlier gneisses have δ^{18}O of +10 to +12.

Pre-Orogenic Plutonic Suites of Eastern Newfoundland

Very few data are currently available from Late Proterozoic suites of the Holyrood and Cross Hills associations; Nd isotopic results are illustrated in Figure 4a at 430 Ma. The Swift Current Granite has ϵ_{Nd} of +1, and the Louil Hills Granite has ϵ_{Nd} of +6, at their time of formation (570 Ma). The Louil Hills signature is close to depleted-mantle evolution, and suggests virtually no input from older crust. The Simmons Brook granodiorite of the Baie D'Espoir area has a strongly negative value of -6 (at 620 Ma) and has a T_{DM} model age of 1730 Ma. If this result is correct, it has important implications, as it seems to indicate a very old component in the Avalon Zone, which is not evidenced by other relevant data (see discussion).

The geochemistry of this unit appears inconsistent with an anatectic model, and it more likely represents mixing of mantle-derived magma and crustal material significantly older than 1700 Ma. Taken at face value, this suggests that there may be rocks of considerable antiquity beneath this region of the Avalon Zone. Oxygen isotopic compositions for the

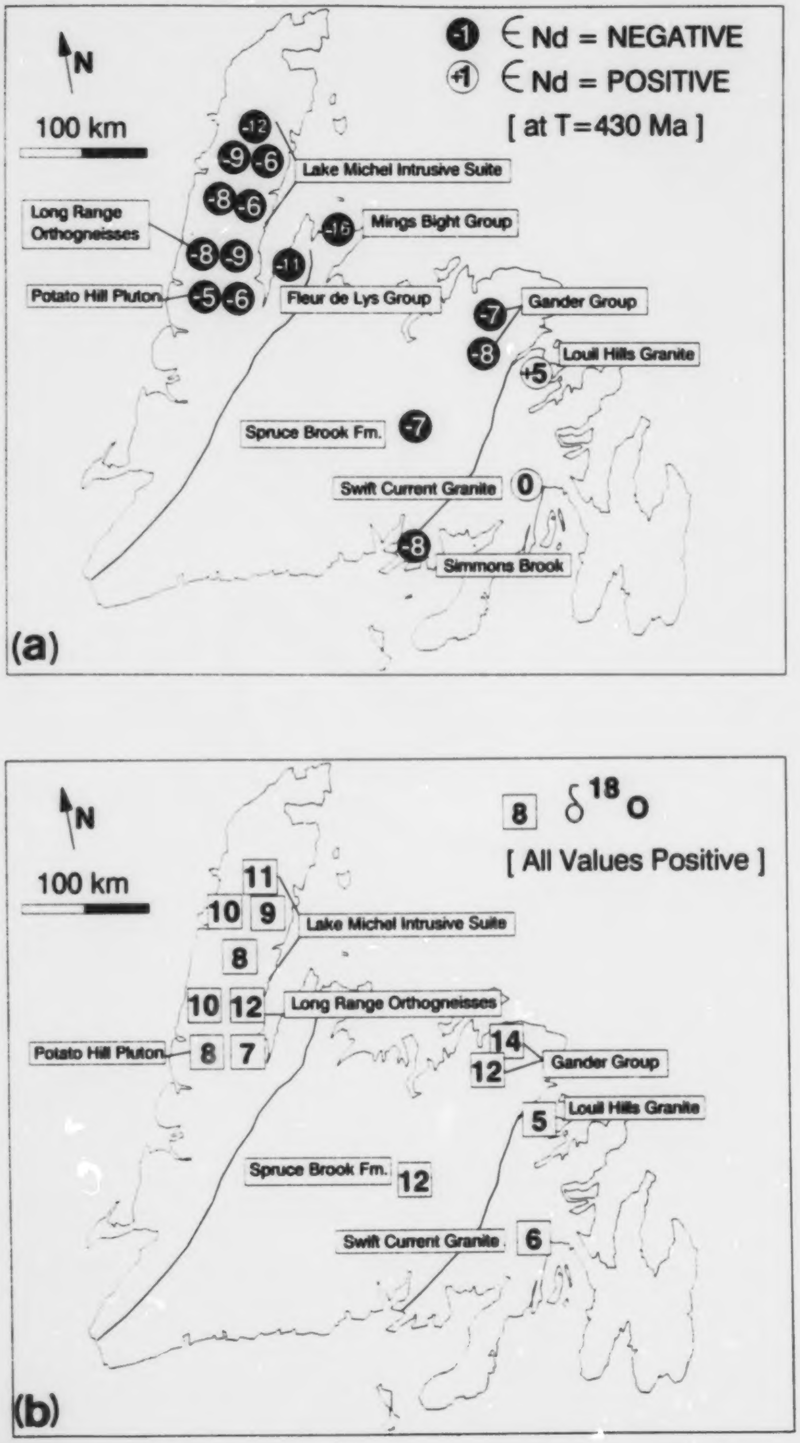
N
100 km
● $\in$Nd = NEGATIVE
+1 $\in$Nd = POSITIVE
[at T=430 Ma]
-12
-9
-6
-8
-6
Lake Michel Intrusive Suite
Long Range Orthogneisses
-16
Mings Bight Group
-8
-9
-11
Potato Hill Pluton
-5
-6
Fleur de Lys Group
-7
Gander Group
-8
Louil Hills Granite
+5
Spruce Brook Fm.
-7
Swift Current Granite
0
-8
Simmons Brook
(a)
N
100 km
8 δ^{18}O
[All Values Positive]
11
10
9
Lake Michel Intrusive Suite
8
10
12
Long Range Orthogneisses
Potato Hill Pluton
8
7
14
12
Gander Group
Louil Hills Granite
5
Spruce Brook Fm.
12
Swift Current Granite
6
(b)

Figure 4. *The Nd and oxygen isotopic characteristics of pre-orogenic plutonic suites and other potential source terranes for Paleozoic granitoid magmas. (a)* $\in_{Nd}$ *of pre-orogenic suites and metasedimentary sequences at T=430 Ma; (b)* $\delta^{18}O$ *of pre-orogenic suites and metasedimentary sequences.*

Louil Hills and Swift Current granites (Figure 4b) are characteristically light and similar to mantle values (+5 to +6).

Recalculation for 430 Ma has only minimal effects (< 2 ∈ units) upon the isotopic features of these pre-orogenic Avalon Zone granites (Figure 4a). The $\in_{Nd}$ value for Simmons Brook at 430 Ma is -8, which is close to the Long Range Inlier values. If the Simmons Brook value is valid, it implies that parts of the Avalonian basement may overlap in isotopic composition with the Laurentian basement, leading to obvious discrimination problems.

Metasedimentary Sequences

Sedimentary rocks have isotopic and geochemical features that reflect their erosional sources; during subsequent metamorphic events, they may be melted to produce 'S-type' granites (Chappell and White, 1974; White and Chappell, 1990). Magmas derived from metasedimentary sources may also contaminate and/or mix with those derived from lower crustal or subcrustal regions. For these reasons, a knowledge of the isotopic features of the main metasedimentary packages in the Central Mobile Belt is very important.

The Gander Group of northeastern Newfoundland and the Spruce Brook Formation, which outcrops in structural windows within the Dunnage Zone, both have $\in_{Nd}$ of about -7 at 430 Ma (Figure 4a), with model ages of 1900 to 1700 Ma. It should be noted that these values represent composite samples, not single locations, and thus represent the bulk features of these sequences. A single zircon and sphene study of a Gander Group psammite (T. Krogh, *in* O'Neill, 1991) gave a range of detrital zircon ages from 2700 Ma to 550 Ma, and sphene ages of around 540 Ma; similar detrital ages of 1600 to 1200 Ma have been reported from the Spruce Brook Formation (T. Krogh, unpublished data). This suggests that the T_{dm} values of 1900 to 1700 Ma reflect simply a mixture of older, but dominantly Proterozoic sources. Oxygen isotopic signatures of Gander Group composites are characteristically heavy, having $\delta^{18}O$ of +10 to +14.

The Fleur de Lys Supergroup, representing the continental rise prism of the ancient Laurentian margin, has ϵ_{Nd} of -11 (at 430 Ma) and model ages of 2000 Ma or older. These sediments thus contain some crustal material that is older than the gneisses sampled in the Long Range Inlier (their presumed source region). This could reflect small quantities of detritus from the central, oldest, part of the Canadian Shield.

The Ming's Bight Group, in the north of the Baie Verte Peninsula, has ϵ_{Nd} of about -16, and a model age of 2400 Ma. These sediments almost certainly contain Archean detritus; although they have been correlated with the Fleur de Lys Supergroup (Hibbard, 1983) they must have had quite different source areas. These results may argue against their correlation, but the limited information on the Fleur de Lys Supergroup may not document internal stratigraphic or geographic variations within this thick sequence. The most important point about this result is that a very ancient potential source or contaminant exists in the eastern Baie Verte Peninsula; a very small amount of Ming's Bight Group entering a rising magma could have dramatic effects on its Nd isotopic signature. This is particularly important in interpreting the rather complex patterns of Paleozoic plutonic rocks in this area (see below).

Further Work

It is clear from the above that our knowledge of the isotopic features of the late Precambrian crust of the Avalon Zone is inadequate. This places restrictions on our interpretation of data from younger suites in the Avalon and Gander zones. Several Avalonian intrusive suites were sampled in 1991, and are currently being analyzed. The present lack of data from late Precambrian metamorphic rocks and plutonic suites on the south coast (Dunning and O'Brien, 1989) and in the eastern Dunnage Zone (Evans *et al.*, 1990) is also of concern, as these represent possible basement to these areas.

Metasedimentary sequences in the Central Mobile Belt are also problematical. On the basis of existing data, they appear to be isotopically distinct, but there are not enough analyses to be certain that there are not equally significant variations *within* these sequences, or if there are systematic variations linked to stratigraphy or geography. It would also be very useful to have a firm picture of the detrital zircon populations in these rocks, preferably in a form that would allow estimates of the proportions of different-aged material to be made. The best method of obtaining the latter would be by ion-microprobe zircon studies, rather than conventional single zircon geochronology.

EARLY OROGENIC PLUTONIC SUITES

Early orogenic plutonic suites for which isotopic data are currently available fall into three clusters (Figures 5a, b).

Notre Dame Bay Area

The Twillingate trondhjemite complex (Williams and Payne, 1975; Payne and Strong, 1979) has ϵ_{Nd} of +3 to +6 at 510 Ma (Figure 5a); however, the majority of the samples lie at the high end of this range. The Burlington granodiorite and the Dunamagon granite of the Baie Verte Peninsula (Hibbard, 1983) have significantly lower ϵ_{Nd} of +1 to +2 and 0, respectively (Figure 5a). Twillingate has $\delta^{18}O$ of +8 to +10 (Figure 5b); no data are available from Burlington or Dunamagon. Recently, Jenner *et al.* (1991) examined trondhjemitic rocks of the Little Port and Bay of Islands complexes in western Newfoundland (Figure 5a). The Little Port Complex has ϵ_{Nd} of -1 to +1 at 505 Ma, which resembles the signature of Burlington. It is interpreted by Jenner *et al.* (1991) as an arc-type magma, with an older component introduced by subduction (see below). Trondhjemites of the slightly younger Bay of Islands Complex have ϵ_{Nd} of +6, and are more akin to the Twillingate data. The pre-Silurian Top Pond Brook tonalite in southwestern Newfoundland has ϵ_{Nd} of around 0.

It is clear that these suites do not contain very large amounts of older crust. This is consistent with their interpretation as products of subduction-related magmatism in an 'oceanic' domain, where older continental crust is not present in direct form (see below). Studies of arc-volcanic rocks in the ocean basins (e.g., White and Patchett, 1984) indicate that most have lower ϵ_{Nd} than the modern depleted-mantle (as represented by MORB), and must therefore include minor older crustal material. The most likely explanation is that continent-derived sediments were subducted into the zones of magma generation. The contrasts in ϵ_{Nd} amongst these suites may reflect contrasts in their positions (relative to continental margins) or in the age of the oceanic crust (and hence the thickness of its sediment cover) that was subducted. However, the probable major strike-slip displacements following closure of Iapetus render any attempt to relate present-day and original positions intrinsically hazardous. Nevertheless, we suspect that systematic geographic isotopic variations lurk within the Ordovician plutonic suites; finding these, and integrating the data with isotopic studies of volcanic suites, are major challenges for future work.

South-Central Newfoundland

The Partridgeberry Hills, Great Burnt Lake and Through Hill granites are now known to be Ordovician (ca. 470 Ma; Colman-Sadd *et al.*, *in press*). The first two have ϵ_{Nd} of -7 and -2 respectively (Figure 5a) and $\delta^{18}O$ of +9 to +11 (Figure 5b). These obviously have radically different origins from the Ordovician suites of Notre Dame Bay. The Partridgeberry Hills granite is within the spectrum of Gander Group metasedimentary rocks (Figure 4a). This, combined with its 'S-type' compositional affinities, indicates that it could have been derived largely by melting of the Gander Group. The high $\delta^{18}O$ values also support this interpretation.

The timing of emplacement of the eastern Dunnage Zone over the Gander Zone has been a subject of much debate, and there is growing evidence that this event occurred during the Ordovician (e.g., Dec and Colman-Sadd, 1990; Colman-Sadd *et al.*, *in press*). Ordovician granites in this area may

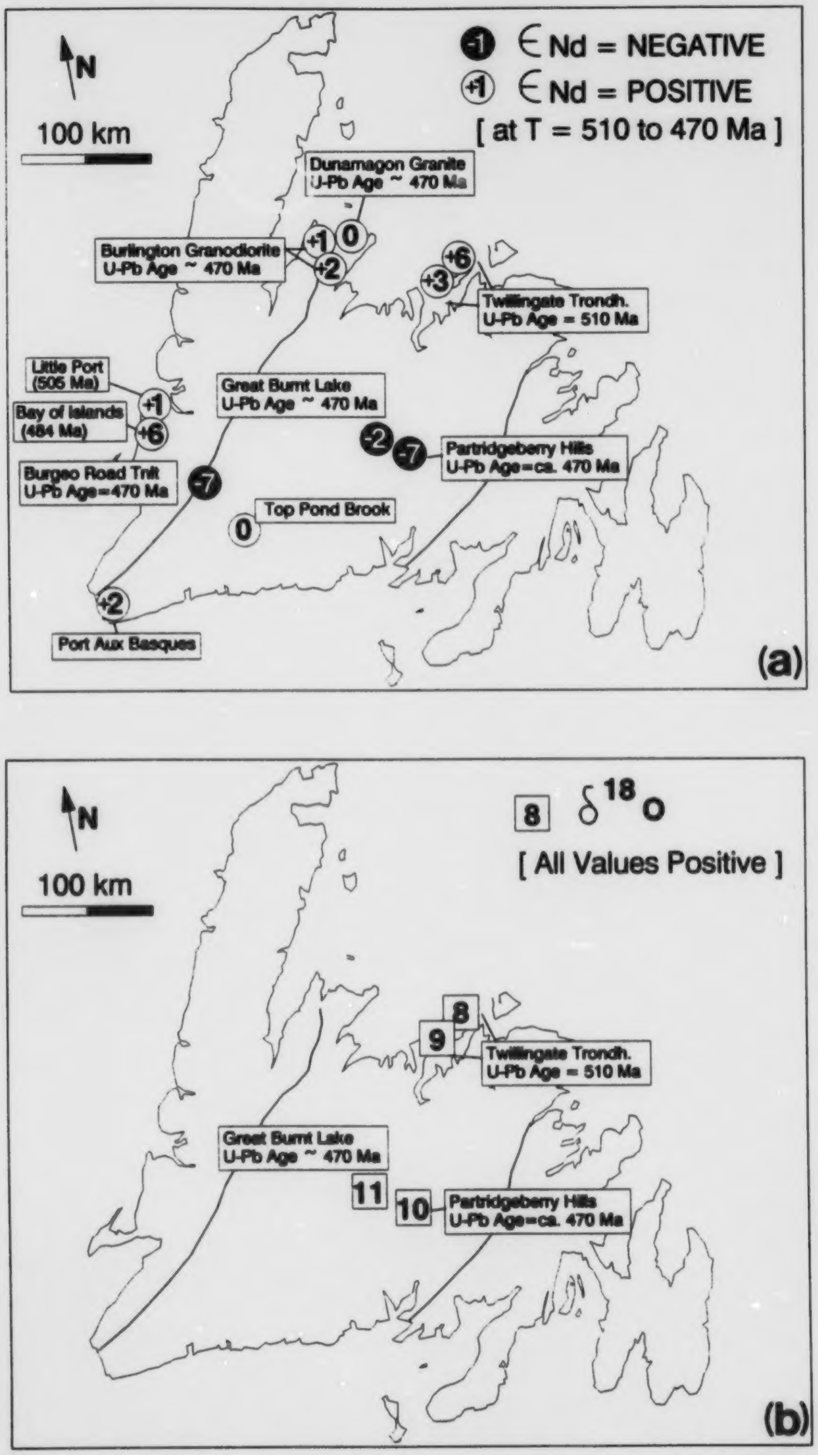
N
100 km
● ϵNd = NEGATIVE
(+1) ϵNd = POSITIVE
[at T = 510 to 470 Ma]
Dunamagon Granite
U-Pb Age ~ 470 Ma
Burlington Granodiorite
U-Pb Age ~ 470 Ma
+1
0
+2
+6
+3
Twillingate Trondh.
U-Pb Age = 510 Ma
Little Port
(505 Ma)
Bay of Islands
(484 Ma)
+1
+6
Great Burnt Lake
U-Pb Age ~ 470 Ma
-2
-7
Partridgeberry Hills
U-Pb Age=ca. 470 Ma
Burgeo Road Tnlt
U-Pb Age=470 Ma
-7
0
Top Pond Brook
+2
Port Aux Basques
(a)
N
100 km
8 δ^{18}O
[All Values Positive]
8
9
Twillingate Trondh.
U-Pb Age = 510 Ma
Great Burnt Lake
U-Pb Age ~ 470 Ma
11
10
Partridgeberry Hills
U-Pb Age=ca. 470 Ma
(b)

Figure 5. *The Nd and oxygen isotopic characteristics of early orogenic plutonic suites. (a)* ϵ_{Nd} *at formation; (b)* $\delta^{18}O$ *variations.*

thus reflect the thermal consequences of accretion within the underlying Gander Zone. Interpretation of the Great Burnt Lake result is more difficult, as this granite could not come entirely from a Gander Group source. One possibility is that its signature instead reflects a largely unexposed basement terrane below the supracrustal cover of the Gander Zone. This option, and the possible nature of this basement, are discussed further below.

Southwestern Newfoundland

Here, Dunning *et al.* (1989) report a U–Pb zircon age of ca. 460 Ma from a foliated tonalite that intrudes ophiolitic remnants. A sample from this unit displays a strongly negative ϵ_{Nd} (-7) at 470 Ma (Figure 5a); this value clearly indicates that the tonalite includes *very significant* amounts of old crust.

There are several ways to interpret this result, and more data are required before any preference can be established. Dunning *et al.* (1989) identified a significant proportion of older 'inherited' zircons, yielding an upper intercept of ca. 1430 Ma. The trace-element signature is strongly depleted in incompatible elements, and seems inconsistent with an origin by anatexis of Grenville Province gneisses or metasedimentary cover. A second alternative is mixing between a tonalitic magma and rocks including a very old component; e.g., something akin to the Ming's Bight Group, which probably contains Archean detritus. A third option is that the sample is a Proterozoic component of the Grenville Province, but this seems at variance with the geochronological evidence. Clearly, more data are required in order to understand this rather puzzling result. Southwestern Newfoundland represents a significant gap in the present isotopic database, and more work is needed in this area, where tonalitic rocks of both Proterozoic and Paleozoic age may be present.

Another potentially interesting result in this area comes from the Port-aux-Basques granite, which has ϵ_{Nd} of +2. The age of this is not documented; van Staal *et al.* (1991) suggest that it has experienced all the deformational events in this area, implying that it may be Ordovician. This is a muscovite-bearing granite with the hallmarks of an anatectic 'S-type' magma, presumably derived from the enveloping paragneisses.

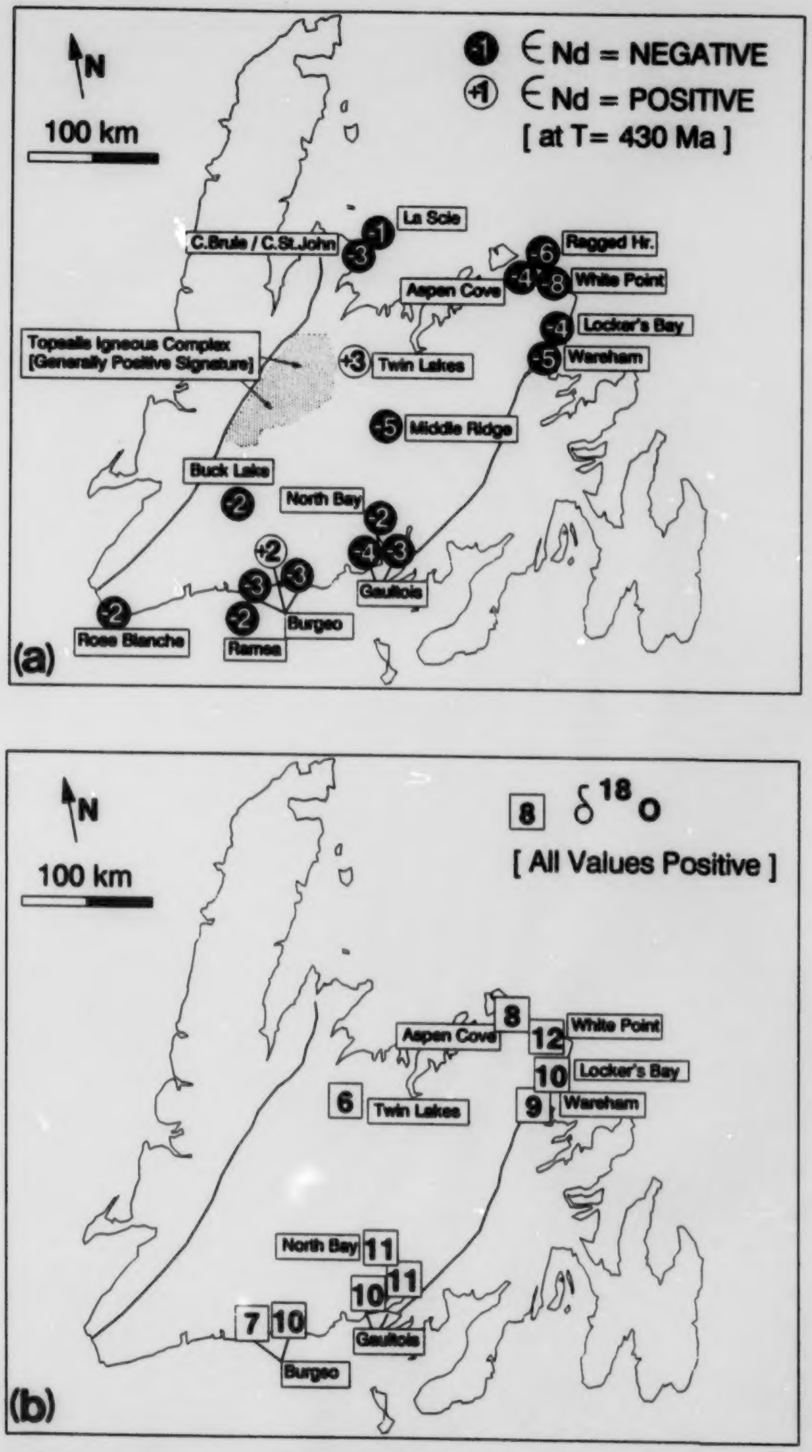
N
100 km
-1 ∈Nd = NEGATIVE
+1 ∈Nd = POSITIVE
[at T= 430 Ma]
La Scie
-1
C.Brule / C.St.John
-3
Ragged Hr.
-6
-4
-8
White Point
Aspen Cove
-4
Locker's Bay
-5
Wareham
Topsails Igneous Complex
[Generally Positive Signature]
+3
Twin Lakes
-5
Middle Ridge
Buck Lake
-2
North Bay
-2
-4
-3
+2
-3
-3
Gaultois
-2
Rose Blanche
-2
Ramea
Burgeo
(a)
N
100 km
8 δ18O
[All Values Positive]
8
Aspen Cove
12
White Point
10
Locker's Bay
9
Wareham
6
Twin Lakes
North Bay
11
11
10
7
10
Gaultois
Burgeo
(b)

Figure 6. *The Nd and oxyten isotopic characteristics of late orogenic plutonic suites. (a)* $\in_{Nd}$ *at formation, calculated for 430 Ma; (b)* $\delta^{18}O$ *variations.*

LATE OROGENIC PLUTONIC SUITES

The Nd and O isotopic data from plutonic suites of Silurian age are illustrated in Figure 6. Note that some intrusions discussed later under 'post-orogenic suites', may also be of Silurian age. For the sake of simplicity, an age of 430 Ma has been used for all $\in_{Nd}$ calculations.

The Gander Zone

With only two exceptions, late orogenic suites from this area have negative $\in_{Nd}$, ranging from -2 to -8 (Figure 6a), and heavy $\delta^{18}O$, generally > +9 (Figure 6b). Muscovite-bearing leucogranite intrusions (e.g., the Middle Ridge, Aspen Cove, Ragged Harbour and White Point granites) have the lowest values (Figure 6a). Their $\in_{Nd}$ ranges from values that resemble the Gander Group (-7) to slightly higher values (-5 to -4). An exception is the North Bay granite in south-central Newfoundland (-2). However, this is a composite batholith (Dickson, 1990), and this single determination may not be fully representative of it.

The negative $\in_{Nd}$ and heavy $\delta^{18}O$ of the Middle Ridge association granites, in conjunction with their peraluminous 'S-type' geochemistry (e.g., Williams *et al.*, 1989; Kerr *et al.*, 1990), support their derivation largely by anatexis of metasedimentary rocks similar to the Gander Group. The closest fit to Gander Group isotopic compositions is shown by the Ragged Harbour and White Point granites of the Carmanville area (Strong and Dickson, 1978; Currie *et al.*, 1980). The latter is a nebulitic granite associated with migmatites; both appear to be products of local anatexis.

The Middle Ridge granite includes a complex, inherited zircon population (Tucker, 1990), which may be as old as Middle Proterozoic (ca. 1600 Ma). This is consistent with derivation from the Gander Group, in which detrital zircons of this general age are abundant (T. Krogh, *in* O'Neill, 1991). However, this example, and other representatives, have slightly higher ϵ_{Nd} than the Gander Group composites. Similar displacement from host metasediments is shown by S-type granites of the Lachlan Fold Belt in Australia (McCulloch and Chappell, 1982), and the South Mountain Batholith of Nova Scotia (Clarke *et al.*, 1988).

As is commonly the case with isotopic data, this can be interpreted in two ways. First, they could be derived from an unexposed metasedimentary package that has a younger provenance than the Gander Group, or include a component from an underlying basement complex having a shorter crustal residence period. Second, they could be mixtures of a mantle-derived magma (with neutral or positive ϵ_{Nd}) and melts generated from metasedimentary rocks. There is no evidence of contemporaneous mafic magmatism in these intrusions, and most are characterized by extremely restricted compositional spectra (Kerr *et al.*, 1990). Thus, the first alternative is preferred. As is discussed below, the identity and nature of a possible basement to the Gander Zone is an important and (as yet) unresolved issue. The Rose Blanche granite in southwestern Newfoundland, grouped with the Middle Ridge association by Williams *et al.* (1989), has ϵ_{Nd} of -2, suggestive of a source significantly different from the Gander Group.

The Burgeo association plutonic suites are characterized by negative ϵ_{Nd} in almost all areas, but typically have ϵ_{Nd} of -3 or higher (Figure 6a), above the upper limit of most Middle Ridge association granites. Indeed, if the Gander Zone is viewed in its entirety, it displays systematic variation in ϵ_{Nd}, from values down to -8 in the northeast, to around -3 on the south coast (Figure 6a). It is not clear if a comparable pattern exists in the relatively sparse oxygen data (Figure 6b). This variation correlates with the distribution of metasedimentary rocks of the Gander Group and its equivalents, which are concentrated in the northeast, but rare or absent on the south coast. The empirical link suggests that the Gander Group supracrustal rocks are in part responsible for these negative signatures, as suggested above. It also supports regional depth zonation, with the south coast representing deeper levels of erosion, where the influence of supracrustal rocks is significantly diminished.

In the Burgeo Intrusive Suite, 5 samples that represent a compositional range from megacrystic biotite–granite to muscovite-bearing leucogranite all have ϵ_{Nd} of around -3. This suggests that these varied components of the suite had similar sources, and may be part of a single magma series. The Gaultois, Ramea and Buck Lake granites have comparable signatures; the slightly more negative character of the Locker's Bay and Wareham granites suggests a greater input from Gander Group materials, consistent with their location. An earlier determination of a ca. 410 Ma muscovite granite sample from the Burgeo Suite yielded a surprisingly juvenile value of +2 (Figure 6a). This result seems at odds with all of the other data from this area, and requires confirmation by re-analysis. There are also tonalitic rocks within the area of the Burgeo Intrusive Suite (Dickson *et al.*, 1989), which have not yet been analyzed. These may provide evidence of a juvenile (subcrustal) component involved in magma generation. Burgeo association suites could therefore be a mixture of three separate components, i.e., a mantle-derived magma, sialic crustal rocks of probable Precambrian age, and metasedimentary rocks with an ancient provenance, particularly in northeastern Newfoundland.

Western Newfoundland

There are presently few data from central and western Newfoundland, although some samples have been analyzed by J. Whalen and G.A. Jenner (unpublished data) and more are currently being processed from 1991 field work by A. Kerr.

On the Baie Verte Peninsula, two volcanic or subvolcanic suites of presumed Silurian age (Cape Brule Porphyry and Cape St. John Group) both yield ϵ_{Nd} of around -3 (Figure 6a). It is hard to interpret these in isolation, but they appear inconsistent with data from the Topsails Complex and post-orogenic intrusive suites in this area, which are generally positive (see below). The proximity of the Mings Bight Group (ϵ_{Nd} = -16 at 430 Ma), raises the possibility of high-level contamination, and their features may not be representative of deep geology. More samples are being analyzed from this area.

The largest Silurian intrusive complex in the Notre Dame Subzone, the Topsails Intrusive Suite, has not been studied as part of this project. However, J. Whalen (personal communication, 1991) reports that many of the A-type granites within the suite display positive ϵ_{Nd} indicative of mantle derivation or only minimal contamination by older crust. Dioritic rocks of the ca. 435 Ma Twin Lakes intrusion also show positive signatures of ca. +3 (Figure 6a) and mantle-like $\delta^{18}O$ values (Figure 6b). Although this would be expected for mafic magmas, the spatially associated Hodges Hill Granite (see below) has almost identical isotopic features. Thus, although the picture is far from complete, the Silurian plutonic suites of the western Dunnage Zone appear to differ radically from those of the Gander Zone, as would also be inferred from the strong contrasts in their elemental geochemistry (e.g., Strong, 1980; Kerr *et al.*, 1990).

POST-OROGENIC PLUTONIC SUITES

The regional Nd and O isotopic variations amongst these suites are illustrated in Figure 7. Note that geochronological studies may force revision of some earlier ages based on Rb–Sr, K–Ar or Ar–Ar data. For the sake of simplicity, an age of 400 Ma has been used in all calculations of ϵ_{Nd}.

Southeastern Newfoundland

The most coherent grouping of Devonian plutons in Newfoundland is the Ackley association of Williams *et al.* (1989). The 14 suites analyzed to date fall into two distinct geographic and isotopic groups (Figure 7).

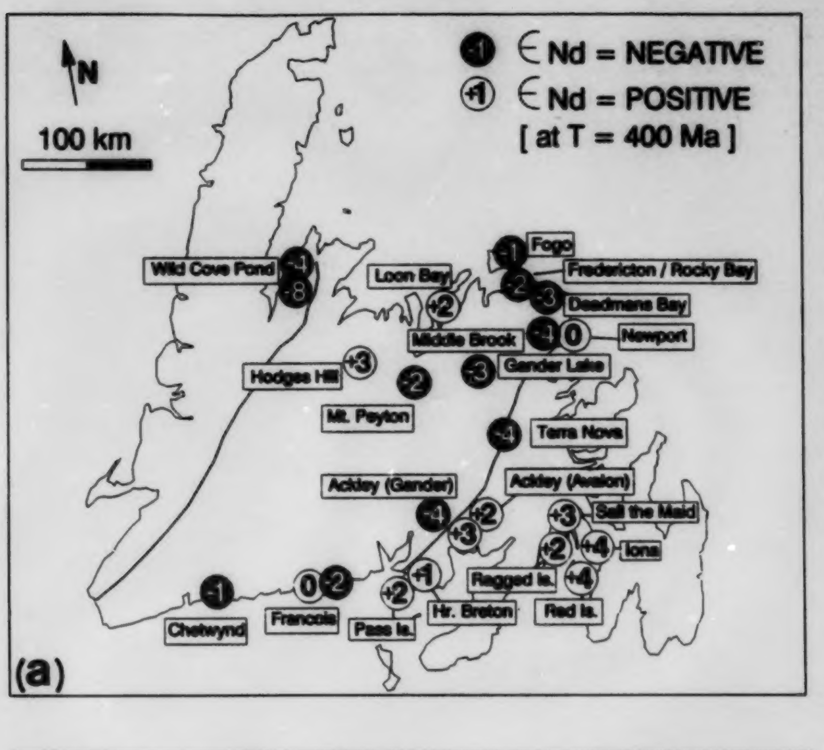
N
100 km
-1 $\in$ Nd = NEGATIVE
+1 $\in$ Nd = POSITIVE
[at T = 400 Ma]
Wild Cove Pond
-1
-8
Loon Bay
+2
Fogo
-1
Fredericton / Rocky Bay
-2
-3
Deadmans Bay
Middle Brook
-4
0
Newport
Hodges Hill
+3
-2
-3
Gander Lake
Mt. Peyton
-4
Terra Nova
Ackley (Gander)
Ackley (Avalon)
-4
+2
+3
+3
Sall the Maid
+2
+4
Iona
Ragged Is.
+4
Red Is.
+1
+2
Hr. Breton
Pass Is.
-1
Chetwynd
0
-2
Francois
(a)
N
100 km
8 $\delta^{18}O$
[All Values Positive]
9
Deadmans Bay
10
Middle Brook
10
Gander Lake
7
Hodges Hill
Mt. Peyton
10
Ackley (Gander)
Ackley (Avalon)
10
7
(b)

Figure 7. *The Nd and oxygen isotopic characteristics of post-orogenic plutonic suites. (a) $\in_{Nd}$ at formation, calculated for 400 Ma; (b) $\delta^{18}O$ variations.*

Within the Avalon Zone, almost all have $\in_{Nd}$ in the range +1 to +4 (Figure 7a); there seems to be an eastward increase in $\in_{Nd}$. Note, however, that the Ragged Islands and Sall the Maid plutons are assumed to be Devonian purely on the basis of lack of deformation, which may not be a fully reliable criterion in the eastern Avalon Zone. These values are not high enough to be purely mantle signatures, but they do suggest the absence of very old crust beneath most of the Avalon Zone. The unusual negative value from the 620 Ma Simmons Brook Granodiorite (Figure 4a; see earlier discussion) is the only countering observation. The Terra Nova Granite has $\in_{Nd}$ of -4 (Figure 7a), which resembles Gander Zone examples. However, this is very close to the inferred Gander–Avalon boundary, and the crust below it may not be part of the Avalon Block. Relatively few oxygen determinations are available, but the Avalon side of the Ackley Granite has mantle-like $\delta^{18}O$ of +7 (Figure 7b).

Within the Gander Zone, these granites are significantly more varied in isotopic composition. As in the case of Silurian suites (Figure 6), they display a regional variation (Figure 7a) from northeast ($\in_{Nd}$ of about -4) to the south coast ($\in_{Nd}$ = -2 to -1).The northeastern values are slightly higher than those of nearby Silurian plutonic suites. Their $\delta^{18}O$ values are also distinctly higher (around +10) than on the Avalon Zone side (Figure 7b). The regional isotopic trends, in our view, are once again linked to the distribution of Gander Group supracrustal rocks, and suggest that a contribution from these materials is *in part* responsible for the negative signatures. However, metasedimentary rocks cannot be entirely responsible, as they are rare on the south coast, where the Nd isotopic signatures of south coast granites remain distinct from those of Avalon Zone examples by 3 to 4 $\in$ units (Figure 7a).

Dunning and O'Brien (1989) present evidence for the existence on the south coast of late Precambrian rocks that have compositional and temporal affinities with the lower-grade rocks of the type Avalon Zone. However, contrasts across the Hermitage Bay Fault at the western unit of the Avalon Zone (Figure 7), and the predominantly negative character of all plutonic suites on the south coast may imply significantly older (unexposed ?) crust with ages up to 1700 to 1300 Ma, or that the identified late Precambrian remnants were themselves derived from older material. Thus, late Precambrian 'basement rocks' on the south coast, although of similar age to the Avalon Zone, may have somewhat different affinities and origins. These problems can be investigated relatively simply by establishing the isotopic features of these enclaves *directly*, which should be a priority for the near future.

The strong contrasts across the Dover–Hermitage Bay fault system support the inferences of Keen *et al.* (1986) and Marillier *et al.* (1989) that this structure is a vertical, crustal-scale break. However, one of the more curious features of the onland transect (Quinlan *et al.*, 1991) is the lack of definition of the Hermitage Bay structure. One possiblity is that the structure may be difficult to image because it is sealed by a 'stitching pluton' similar to the Ackley Granite, several kilometres below the surface. On the basis of the isotopic contrasts displayed in Figure 7, we suggest that the Dover–Hermitage Bay fault *does* represent a major break, although it may perhaps only separate different segments of a composite 'Gondwanan' Precambrian terrane.

The Ackley Granite, which intrudes the Dover–Hermitage Bay fault and 'stitches' the Gander and Avalon zones, exhibits internal Nd isotopic contrasts, from -4 on the Gander side to +3 on the Avalon side, and also shows a contrast in oxygen isotopic compositions, as noted by Tuach *et al.* (1988). This large high-silica magma chamber has inherited and somehow retained the features of two separate source terranes. These variations suggest that it is a composite pluton or, alternatively, they place limits on the scale of any internal homogenization processes (e.g., convection) that operated within a large magma chamber of the type envisaged by Tuach *et al.* (1986).

The François Granite (ca. 380 Ma; Tucker, 1990) also displays interesting isotopic variations. The ϵ_{Nd} value ranges from -2 in the oldest phases to around 0 in the youngest component, an evolved U-rich granite that forms the core of one of two ring-complexes. This suggests significant changes in the petrogenesis of magmas during the evolution of the complex, and stresses the importance of analyzing several samples from lithologically varied suites.

The other exception to the general negative pattern of Gander Zone intrusions is the Newport Granite of Bonavista Bay, whose ϵ_{Nd} averages around 0. This is also a young granite, and lies immediately west of the Gander–Avalon zone boundary.

Western Newfoundland (Humber Zone)

The only analyses to date are from the Wild Cove Pond Intrusive Suite of the western Baie Verte Peninsula (Hibbard, 1983). This composite batholith includes dioritic rocks and K-feldspar megacrystic granites that (at least superficially) resemble those of the northeast Gander Zone. Subordinate muscovite-bearing leucogranites, locally foliated, are also present. Nd isotopic compositions are related to lithology; dominant megacrystic biotite-granites have ϵ_{Nd} of about -4, whereas foliated leucogranites have lower values of around -8. The latter is similar to the composition of Grenville material at about 400 Ma, but slightly higher than typical Fleur de Lys Supergroup (-11 at 400 Ma; Figure 4d). The presence of gabbroic and dioritic rocks in the Wild Cove Pond Suite suggests that mantle-derived magmas may be implicated in its generation, and it seems reasonable to suggest mixing of this material with Grenville crust and/or Fleur de Lys metasediments. The leucogranites perhaps represent the crustal end-member component of the mixture.

Central Newfoundland

Small amounts of interesting data are available from central Newfoundland. In the Notre Dame Subzone, the Hodges Hill granite (ca. 410 Ma; Rb–Sr; B. Fryer, unpublished data) and the Loon Bay Granodiorite (408 ± 2 Ma; U–Pb, Elliot *et al.*, 1990) both have ϵ_{Nd} of +2 to +3. Neither result seems consistent with the presence of significant amounts of Grenville basement beneath this area, as originally suggested by the offshore seismic-reflection lines (Keen *et al.*, 1986). Coupled with data from the Twin Lakes Diorites (Figure 6a) and the Topsails Intrusive Suite (J. Whalen, personal communication, 1991), these may argue for the presence of a relatively juvenile, possibly early Paleozoic crustal block in this area, rather than the Grenville crust suggested previously. This is a critical area for future data acquisition.

In contrast, plutonic suites that are close to the Dunnage–Gander boundary (e.g., Mt. Peyton; Fredericton and Rocky Bay plutons) have ϵ_{Nd} of around -2, similar to signatures from those parts of the Gander Zone where metasedimentary sequences are poorly developed. The Fogo granite has a rather indeterminate signature of around -1; the abundant dioritic rocks at Fogo have not been analyzed yet.

On a first-order basis, the boundary between the positive and negative domains appears to be roughly coincident with the Red Indian Line, defined by Williams *et al.* (1988) as the boundary between the Notre Dame and Exploits subzones of the Dunnage Zone. There also appear to be contrasts in elemental geochemistry between the post-orogenic intrusive suites on either side of this line (Kerr *et al.*, 1990), but the data are not sufficient to draw firm conclusions. The correspondance with the Red Indian Line is not exact, as the Loon Bay Granodiorite actually lies within the Exploits Subzone.

The isotopic data do, however, suggest some fundamental significance for the Red Indian Line, which is roughly coincident with the western limit of typical Gander Zone isotopic signatures. The nature of the material to the northwest of the Red Indian Line is conjectural; it is possible that it represents a zone of relatively juvenile sialic crust, perhaps representing material generated at arcs developed at various times and places within Iapetus.

CONCLUDING REMARKS

It is clear that we do not yet fully comprehend the details of the isotopic patterns discussed in this report, and that many gaps remain in our knowledge. However, several points emerge; some have fundamental geological significance, whereas others point mostly at directions for continued research in the final year of the project.

1. The features of the Grenville Block are relatively well-established, although some data from the Indian Head Complex in southwestern Newfoundland are required.

2. The features of the Avalon Block are very poorly known. The present Nd and O isotope data do not characterize the late Precambrian crust of this area with enough reliability to allow it to be recognized elsewhere. It is also vital that late Precambrian rocks from the south coast and central Newfoundland be characterized, as these represent potential basement to the Gander Zone. The possibility of a late Precambrian 'Gondwanan' basement in central Newfoundland renders studies of the 'type' Avalon Zone very important.

3. Both the Gander Group–Spruce Brook formation and the Fleur de Lys Supergroup have T model ages that are older than the Avalon and Grenville blocks, respectively. However, the first observation is tentative in view of our poor isotopic knowledge of Avalonian rocks. The Ming's Bight Group has a very ancient provenance; this material could have a powerful impact on any magma that encounters or digests it. In general, there is a requirement for more systematic studies of these sequences to establish if there are systematic stratigraphic and/or geographic isotopic variations within them.

4. Isotopic variations amongst Ordovician trondhjemite–tonalite–granodioritic suites may be related to the original locations of different arc systems relative to North America or Gondwanaland and/or their age. Ordovician 'S-type' granites in south-central Newfoundland record metamorphism and anatexis of the Gander Group, probably related to the accretion of the eastern Dunnage Zone. The isotopic features of probable Ordovician suites in the southwest Dunnage Zone are not well established.

5. The signatures of Silurian plutonic suites in the Gander Zone are consistent with the concept of a Late Proterozoic basement overlain by a supracrustal package derived from significantly more ancient Early and Middle Proterozoic sources. This is in accordance with geochronological data, but the relationship between this 'basement' and the type Avalon Zone may not be simple and direct. In the northwestern Dunnage Zone, it appears that most Silurian Suites had relatively juvenile characteristics; these argue against the presence of Grenvillian basement rocks, and perhaps indicate a discrete lower crustal block in this region, possibly representing arc systems accreted to North America.

6. Devonian plutonic suites show strong contrasts across the Gander–Avalon boundary. These isotopic contrasts between the Avalon Zone and Gander Zone plutons persist on the south coast, where metasedimentary rocks cannot be implicated for negative signatures within the latter. The Dover–Hermitage Bay fault system is thus an important isotopic break, despite problems in imaging it in the onland seismic data, and Upper Proterozoic basement west of it may not be *exactly* correlative with the type Avalon Zone, although the two could be parts of a single composite terrane. In central Newfoundland, plutons northwest of the approximate position of the Red Indian Line display relatively juvenile signatures compared to those to the southeast.

7. The Nd and O isotopic data should be augmented with information from other isotopic systems. On the basis of studies elsewhere (e.g., Ayuso, 1986; Ayuso and Bevier, 1991), whole-rock and/or feldspar Pb-isotopic studies are the most promising, as these have defined major crustal blocks and boundaries in a manner similar to that demonstrated here. In combination, the two methods are more powerful than each in isolation, and are more likely to provide unique solutions.

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GEOLOGY OF MARMORIZED, LOWER PALEOZOIC, PLATFORMAL CARBONATE ROCKS, 'PYE'S RIDGE', DEER LAKE

I. Knight
Newfoundland Mapping Section

ABSTRACT

'Pye's ridge' is a northeast-trending hill, 300 m high, just to the north of Deer Lake. The hill, which is underlain by a succession of dolomitic and calcareous marble, has high potential as a source of dimension stone. The polydeformed succession features metamorphosed Middle Cambrian to Middle Ordovician carbonates similiar to those mapped in the adjacent areas of the Pasadena map sheet. The succession includes rocks correlated with the Hawke Bay Formation, Reluctant Head Formation, and the Port au Port, St George, Table Head and Goose Tickle groups. The marbles are juxtaposed structurally against psammitic and pelitic rocks of the Old Man Pond Allochthon to the northwest, and are overlain, unconformably, by flat-lying Carboniferous sediments in the south and, locally, elsewhere.

The rocks of 'Pye's ridge' are polydeformed and are probably part of a duplex. The three main phases of deformation include 1) northwest-verging thrusting, 2) northeast-trending, southeast-verging folding and 3) northwest-verging normal extension. These phases all predate the formation of Carboniferous cave deposits (red sediment and speleothems) along the northwestern terrain boundary and D_2 *faults. Locally, reactivation of faults has deformed the speleothems; C–S fabrics indicate sinistral movement. Rare kink bands may correlate with this late or post-Carboniferous movement.*

The marble of 'Pye's ridge' include several dimension-stone prospects. Varieties range from white (bianco) *to grey* (bardiglio) *to some pink and include a number of textures that range through* arabescato, venato, nuvolato, striato *and* scuro. *One variety of* bardiglio *resembles* b. imperial.

INTRODUCTION

'Pye's ridge' (informal name) (Figures 1 and 2) is a northeast-trending, forested ridge that occurs just to the northwest of the town of Deer Lake, and north of the lake of the same name. The ridge, which is 6 km long by 3 km wide rises to more than 330 m (950 feet). It is the location of an extensive and quite well-exposed succession of lower Paleozoic marble. The marble lies southeast of the Old Man Pond Allochthon, from which they are seperated by the 'Helen's brook' fault that follows a stream valley, informally named by Gillespie (1983) 'Helen's brook'. The marble is currently covered by a claim block operated by Len Pye. The claim block includes a number of interesting marble prospects that were recently evaluted for Technostone, an Italian engineering and dimension-stone company.

A mapping project was initiated by the Geological Survey Branch of the Department of Mines and Energy to delineate the prospective economic marble units and study the structure of the ridge. The ridge contains four known marble prospects, two white, one ivory and one grey. The white marble was originally thought to be unconnected stratigraphically but mapping indicated that the two prospects occur in the lower part of the stratigraphy, repeated by a thrust. The ivory and grey prospects occur stratigraphically higher. Several other marble targets are outlined in this report.

The rocks are deformed by at least three phases of deformation; an early phase of northwest thrusting, a phase of southeast-verging recumbent folding, possibly related to thrusting, and a phase of down-dip northwestward extension. The latter is related to normal-sense fault reactivation along the structural boundary betweeen the marble to the southeast and the Old Man Pond Allochthon to the northwest.

It is possible to define a stratigraphy within the marble that compares with that in rocks of the Goose Arm–Old Man Pond area to the west (Knight and Boyce, 1991) (Figure 3). The rocks are the equivalent of the Reluctant Head Formation, Port au Port Group (Late Cambrian Petit Jardin and Berry Head formations) and the Ordovician St. George and Table Head groups. The lower white marble prospects occur in the Berry Head formation; the stylolitic grey marble occupies a large area within the upper half of the Catoche Formation

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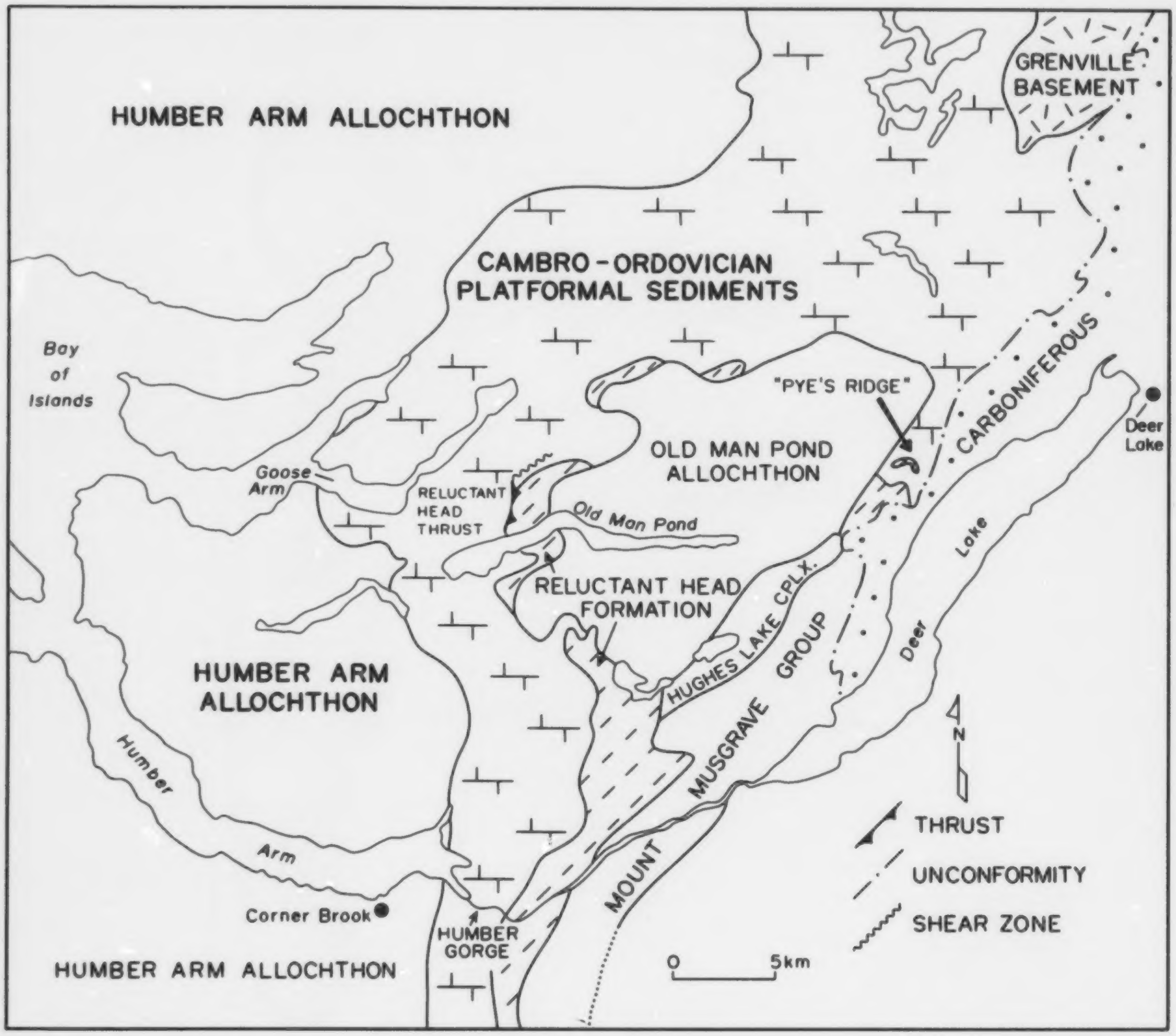

HUMBER ARM ALLOCHTHON
GRENVILLE BASEMENT
CAMBRO-ORDOVICIAN PLATFORMAL SEDIMENTS
Bay of Islands
Goose Arm
RELUCTANT HEAD THRUST
"PYE'S RIDGE"
OLD MAN POND ALLOCHTHON
Old Man Pond
CARBONIFEROUS
Deer Lake
RELUCTANT HEAD FORMATION
HUGHES LAKE CPLX.
MOUNT MUSGRAVE GROUP
Deer Lake
HUMBER ARM ALLOCHTHON
Humber Arm
N
THRUST
UNCONFORMITY
SHEAR ZONE
Corner Brook
HUMBER GORGE
0 5km
HUMBER ARM ALLOCHTHON

Figure 1. *Geology of the Humber Arm–Deer Lake area showing the main geological units and the location of 'Pye's ridge'. Modified from Williams and Cawood (1989).*

of the St. George Group; the ivory marble is the Costa Bay member of the Catoche Formation.

STRATIGRAPHY

CAMBRIAN

Unit 1: Pelite–Psammite

The structurally lowest and presumably oldest rocks exposed in the area consist of green pelites having thin leaves of quartz psammite (Plate 1). They occupy only a few small outcrops that include in two instances the cores of eye folds in lowland to the southwest of the ridge (Eighth Brook area). The unit is probably a distal siliciclastic shelf-facies equivalent of the Hawke Bay Formation and hence of late Early Cambrian to early Middle Cambrian age.

Unit 2: Ribbon Limestone–Phyllite

Silvery-grey to dark-grey phyllites and blue-grey, thin-bedded, dolomitic ribbon limestones overlie the green pelitic unit (Plate 2). It is widely developed and occupies the foot of the slope at the southwest end of 'Pye's ridge' and much of the land around Eighth Brook. It also occurs in a narrow zone trending east-west across the middle of 'Pye's ridge' where it forms the hanging wall sole of the 'Pye's ridge' thrust. The unit is lithologically comparable to the Middle Cambrian Reluctant Head Formation that occurs extensively in the Reluctant Head Thrust at the west end of Old Man Pond (Knight and Boyce, 1991).

Although the unit is strongly deformed internally, the overall upward trend appears to be one of phyllite to carbonate. Silvery-grey and dark-grey phyllites form beds up to 1 m thick in lower parts of the formation and are intercalated with phyllite beds up to 40 cm thick that include thin beds and ?laminations of limestone. Locally, some limestone beds are grainy and others comprise intraformational rudstone. Quartz–calcite–siderite veins are common. In the middle of the formation, the phyllite and limestone occur in roughly equal proportions to each other, forming units up to 1 m or more of interbedded thin beds of ribbon limestone and phyllite. Beds of thinly stratified grainstone, intraformational lime conglomerate and dolostone, reaching 30 to 40 cm thickness also occur. Some grainstone beds are graded with crosslamination at the top. Toward the top of the formation, phyllite is mostly restricted to partings within very compact, thin-bedded limestone. Some of the limestone beds possess a nodular texture. A transition into the overlying dolomite unit is marked by interbeds of ribbon limestone and argillaceous dolostone having limestone lumps, dolostone and intraformational limestone conglomerate. At the same contact within the upper thrust slice, ribbon limestone is dolomitized between thick dolostone beds.

The true thickness of the formation cannot be estimated although structural cross-sections indicate a thickness of approximately 250 to 350 m.

Unit 3: Dolomite Marble

The lower dolostone unit generally comprises thick beds of pale, buff- to yellow-weathering, finely crystalline, structureless, sparcely stylolitic, pale-grey dolostone. Locally, crossbeds and crosslamination are discernable with carbonate grains (now recrystallized) visible. Some beds exhibit thin stratification and locally, some brown-weathering, thinly stratified argillaceous dolostone occurs. Rounded, coarse-grained quartz sand occurs in the middle part of the dolostone, and the upper part of the dolostone unit (subunit 3a) is marked by abundant white to black chert nodules and irregular to folded quartz veins. Toward the top, the dolostones are cleaner and include brown-weathering, massive, cream dolomite marble of fine, massive texture. These characteristics suggest that the dolostone consists of strata equivalent of both the Petit Jardin and Berry Head formations. The latter comprises the cherty dolostones which, in sections at Goose Arm and in the Reluctant Head thrust slice, occupy the lower two-thirds of the formation.

The thickness of the dolomite marble is estimated structurally to range between 250 and 500 m.

Unit 4: Lower Limestone–Dolomite Marble

This unit is characterized by white, ivory, very pale blue-grey, very pale-grey to blotchy off-white and pale-grey limestone marble, yellow-weathering dolostone and carbonate conglomerate or breccia; pink colouration is also locally developed. The unit is extensively developed at the top of the lower thrust slice and is repeated in the upper thrust slice.

The succession from the base upward is as follows–1) interbedded white marble and dolostone; the marble frequently passes laterally into a pure, creamy, massive dolomite marble; 2) limestone conglomerate–breccia (Plate 3) comprising both marble and lesser dolostone blocks in a fine-grained, phyllitic carbonate matrix (1 and 2 combine to form subunit 4a); 3) dominantly white to ivory marble with thin dolostone beds; and 4) off-white, pale blue-grey, pale-grey and some dark-grey marble and interbedded dolostones.

Within this succession, the sedimentary fabrics still preserved include stromatolite mounds, crossbedding, burrow fabrics, parted carbonate, stylolitic bedding, thin stratification and lamination including mudcracks in dolostone beds. White stylolitic marble having phyllitic partings are probably recrystallized peloidal grainstones, which are commonly pale-coloured in less-deformed thrust slices in the Goose Arm area to the west. These characterisistics indicate that the succession is upright. Blotchy grey and off-white marble are probably recrystallized crypt-microbial mounds. Some thicker dolostone beds include interstratification of lamination and intraclastic grainstone–rudstone. These sedimentary features and the interbedded nature of the marble and dolostone, plus the stratigraphic position of the unit above massive dolostones, suggest that the unit is equivalent to the transition at the top of the Berry Head formation mapped at Goose Arm (Knight and Boyce, 1991). The unit attains a thickness of about 200 m.

The limestone conglomerate–breccia (Plate 3) (part of subunit 4a) comprises white marble and yellow-weathering, cream dolomite marble blocks that are mostly massive but also include stylolites and thin stratification. The matrix is dolomitic containing minor white and green micas. Although subunit 4a reaches a thickness of 30 to 50 m, the conglomerate is less extensive and perhaps no thicker than 5 to 10 m. Also, it is not always developed, and locally, in the upper thrust slice, as in the 330 m hill, it is associated with only very thin white marble, and appears instead to be associated with grey and some off-white marble interbedded with dolostone. The explanation for this variation may reflect the origin of the conglomerates. Although they may be channelized resedimented gravity or waterlain deposits, the poor sorting, uneven distribution and varied stratigraphic association suggest that they are metamorphosed interstratal collapse breccias, like those that are common in the autochthonous sections of western Newfoundland (Knight, 1978; Knight and Cawood, 1991).

ORDOVICIAN

Ordovician strata are typified by a thick succession of dominantly dark-grey to light-grey lime marbles having various proportions of dolomite marble overlain by a unit of slate and carbonate. Six rock units are delineated: lower dark-grey marble; varicoloured marble; middle dark-grey marble; ivory marble; upper dark-grey marble, and dark-grey phyllite with marble. This succession is correlated with the carbonates of the St. George and Table Head groups and the flysch of the Goose Tickle Group.

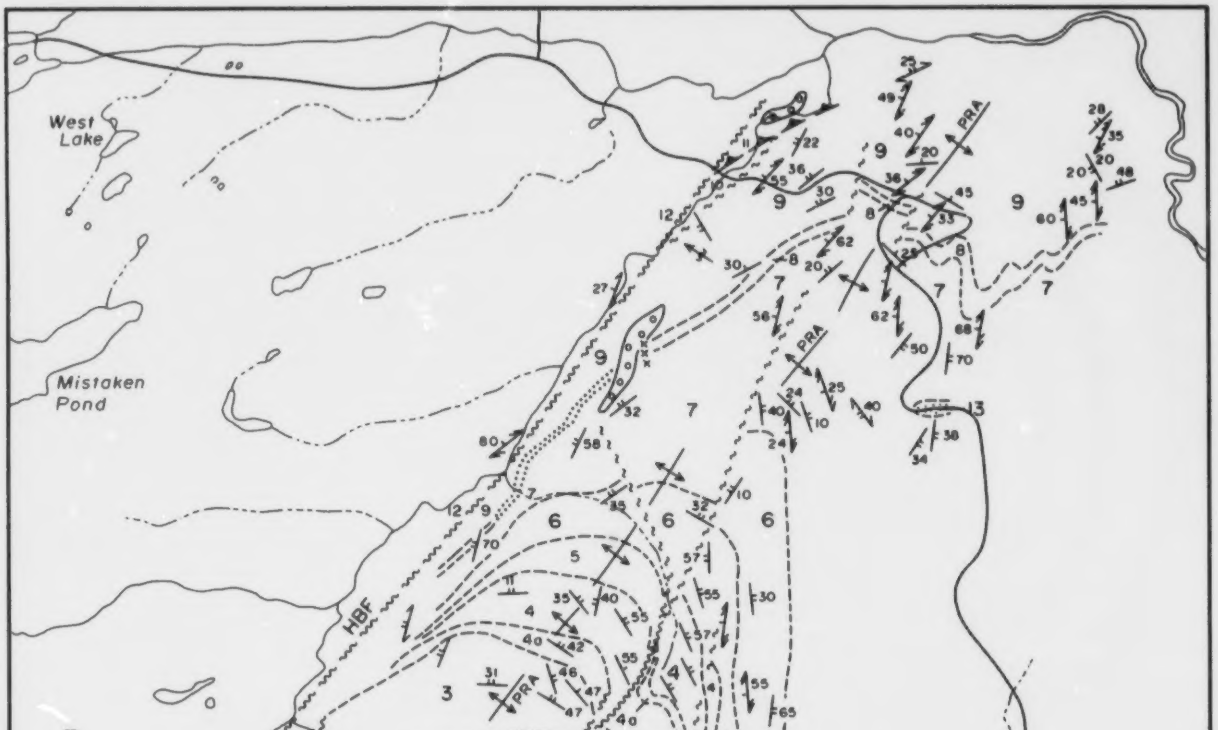

West Lake
Mistaken Pond
HBF
PRA
PRT
Deer Lake
Eighth Brook
N
0
1km

Figure 2. *Geological map of 'Pye's ridge' and adjoining areas.*

LEGEND (for Figure 2)

CARBONIFEROUS

13 *Grey conglomerate–breccia, red conglomerate, sandstone and siltstone*

?PRECAMBRIAN–CAMBRIAN

OLD MAN POND ALLOCHTHON

12 *Psammite and pelite*

?ORDOVICIAN

11 Melange

MIDDLE ORDOVICIAN

GOOSE TICKLE GROUP

10 *Dark-grey phyllite with marble; grey phyllite, thin-bedded and conglomeratic limestone and dolostone*

TABLE HEAD GROUP

9 UPPER DARK GREY MARBLE *Thick-bedded, massive dark-grey to grey limestone; cyclic with fenestral fabrics and rare dolostone at the base*

LOWER ORDOVICIAN

ST. GEORGE GROUP

8 IVORY MARBLE *Ivory- to white-coloured, thinly bedded marbles having green, dark-grey and dolomitic, stylolitic partings; minor dolostone beds at the base*

7 MIDDLE DARK GREY MARBLE *Burrow-dolomite-mottled, thick bedded and stylolitic*

6 VARICOLOURED MARBLE *Vari-grey to white, burrow-mottled, mounds, dololaminite*

5 LOWER DARK-GREY MARBLE *Thick-bedded, burrow-mottled, mounds*

CAMBRIAN

PORT AU PORT GROUP

4 LOWER LIMESTONE–DOLOMITE MARBLE *White, blue to light-grey marble, dololaminite, mounds, crossbedded;* **4a**, *conglomeratic*

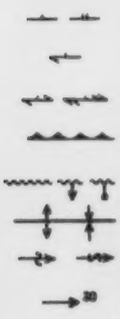
3 DOLOMITE MARBLE *Light-grey to cream, massive, thick bedded;* **3a**, *cherty in upper part*

2 RIBBON LIMESTONE–PHYLLITE *Blue-grey, thin-bedded marble having phyllite and dolomitic beds and partings; some grainy and conglomeratic beds*

LABRADOR GROUP?

1 PELITE AND PSAMMITE *Green-grey to grey pelite having white psammite laminae*

SYMBOLS



Bedding, tops known, unknown

Co-planar bedding-cleavage; schistosity

Cleavage, D_1, D_2

Thrust fault

High-angle fault, dip, downthrow

Fold axes, anticline, syncline

Minor folds, plunge, upright to inclined axis, recumbent to reclined

Lineation, plunge

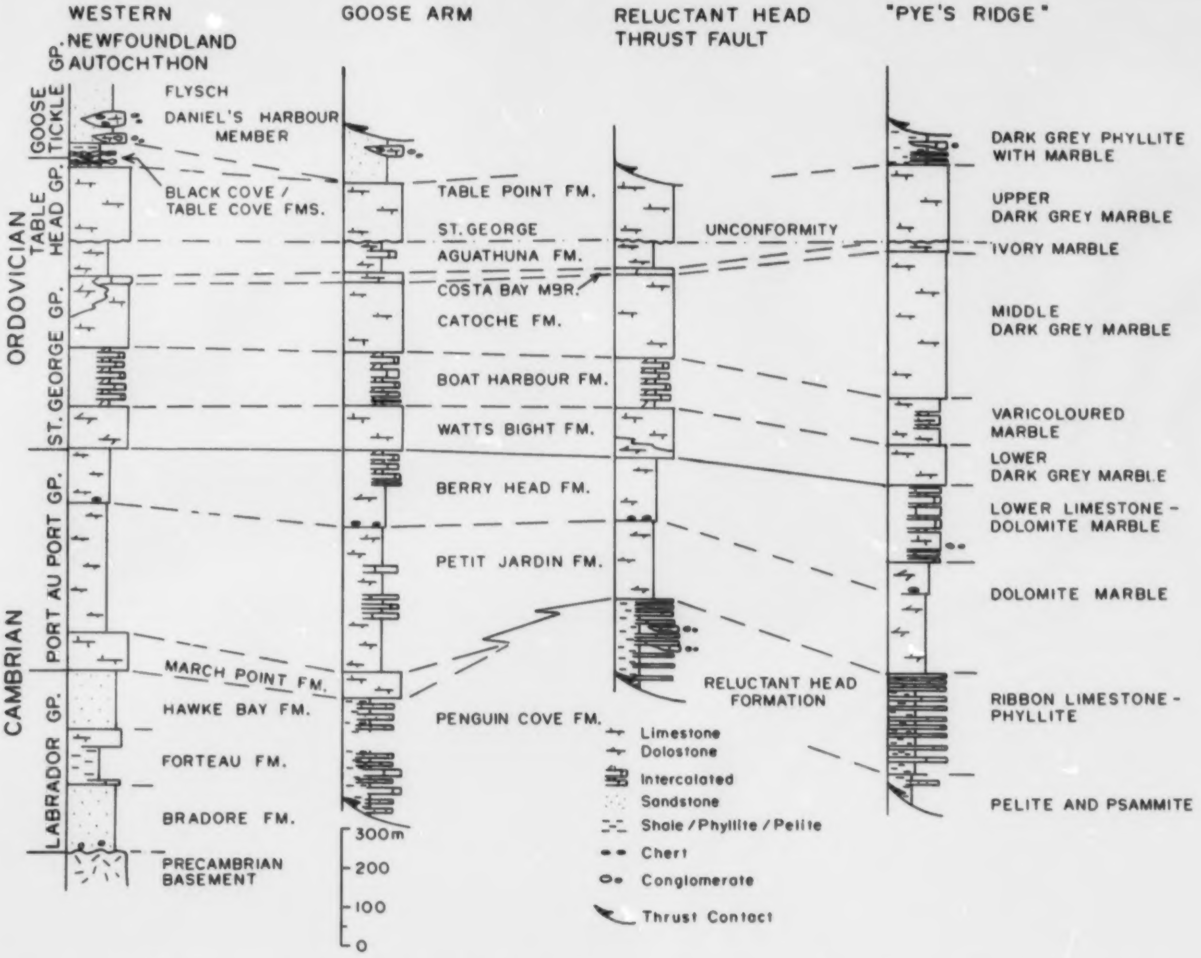

WESTERN NEWFOUNDLAND AUTOCHTHON
GOOSE ARM
RELUCTANT HEAD THRUST FAULT
"PYE'S RIDGE"
ORDOVICIAN
CAMBRIAN
GOOSE TICKLE GP.
TABLE HEAD GP.
ST. GEORGE GP.
PORT AU PORT GP.
LABRADOR GP.
FLYSCH
DANIEL'S HARBOUR MEMBER
BLACK COVE / TABLE COVE FMS.
TABLE POINT FM.
ST. GEORGE
UNCONFORMITY
AGUATHUNA FM.
COSTA BAY MBR.
CATOCHE FM.
BOAT HARBOUR FM.
WATTS BIGHT FM.
BERRY HEAD FM.
PETIT JARDIN FM.
MARCH POINT FM.
HAWKE BAY FM.
FORTEAU FM.
BRADORE FM.
PRECAMBRIAN BASEMENT
PENGUIN COVE FM.
RELUCTANT HEAD FORMATION
DARK GREY PHYLLITE WITH MARBLE
UPPER DARK GREY MARBLE
IVORY MARBLE
MIDDLE DARK GREY MARBLE
VARICOLOURED MARBLE
LOWER DARK GREY MARBLE
LOWER LIMESTONE - DOLOMITE MARBLE
DOLOMITE MARBLE
RIBBON LIMESTONE - PHYLLITE
PELITE AND PSAMMITE
300m
200
100
0
Limestone
Dolostone
Intercalated
Sandstone
Shale / Phyllite / Pelite
Chert
Conglomerate
Thrust Contact

Figure 3. *Comparative stratigraphy of 'Pye's ridge' with that of western Newfoundland autochthon, and the parautochthonous successions of the Goose Arm area and Reluctant Head thrust, Old Man Pond. Based upon Knight (1991),* this report, *unpublished data and Knight and Boyce, (1991).*

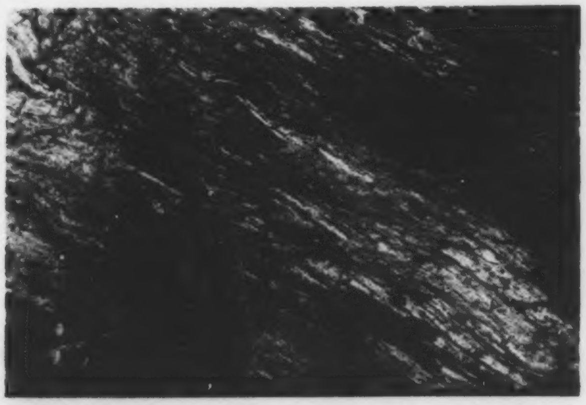


Plate 1. *Intensely deformed pelite and psammite (Unit 1) with rootless psammite folia flattened and folded parallel to the schistosity. Southwest end of 'Pye's ridge'.*

Unit 5: Lower Dark-Grey Marble

Unit 5 consists of approximately 125 m of thick-bedded dark-grey and grey marble interbedded with thin dolomite beds. The marble is characterized by mottle fabrics accentuated by dolomite. These indicate that the original rocks were bioturbated limestone (Plate 4) and microbial mounds. This unit is correlated with the Watts Bight Formation of the St. George Group.

Unit 6: Varicoloured Marble

This is a poorly exposed, 150-m-thick unit of varicoloured lime marble and rare dolomite. Most of the unit is medium to light-grey having whitish streaks and dolomite partings. Original fabrics are obliterated by metamorphism. The unit is correlated with the Boat Harbour Formation of the St.George Group.

Plate 2. *Banded limestone of Unit 2 with phyllitic and dolomitic bands. A thicker limestone layer is boudinaged and the trace of* S_2 *cleavage can be seen dipping steeply to the left. Hanging wall of the 'Pye's ridge' thrust near location 7 (Figure 4).*

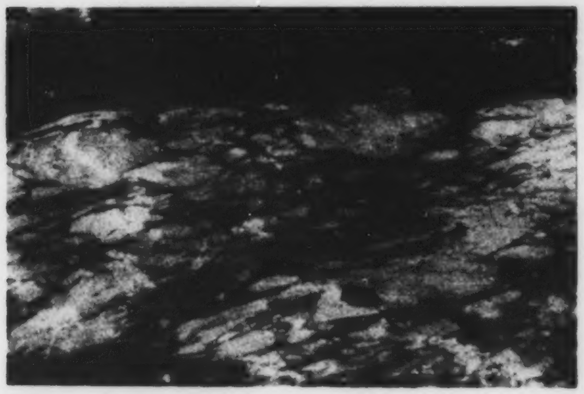


Plate 3. *Limestone conglomerate* (arabescato) *in the lower part of Unit 4, lower thrust slice, location 1 (Figure 4).*

Plate 4. *Dark-grey marble showing inhomogeneous deformation. Dolomitized burrows are present in the less-deformed bed (upper right), which structurally overlies a strongly flattened marble (lower dark-grey marble, upper thrust slice).*

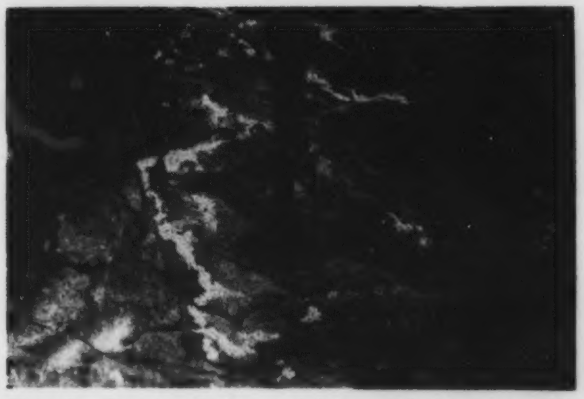


Plate 5. *Grey to off-white, stylolitic marble from the middle dark-grey marble unit on the south side of 'Pye's ridge'. Location 5, Figure 4; photograph courtesy of Jamie Meyer.*

Unit 7: Middle Dark-Grey Marble

This extensively developed unit underlies most of the middle part of the ridge. The marble has dolomitized burrow fabrics and bedding-parallel dolomite seams. Nodular, dolomitic blue-grey marble and blotchy grey to light-grey marble, originally microbial mounds possibly, are also common. Black chert nodules are also present, locally, in the middle part of the formation. Stylolitic grey marble (Plate 5) that occurs in the upper part of the unit underlies an extensive part of the southern slope of the ridge. The stylolites are dark-grey to black. The unit attains a thickness of at least 375 m. This is at least twice as thick as the equivalent unit elsewhere in western Newfoundland.

Unit 8: Ivory Marble

This unit, approximately 20 m thick, is ivory to white with green stylolitic seams and partings reflecting the presence of micas. Off-white to white marble having a dark-grey to black stylolitic seam marks the unit as it is traced to the east. Colour locally varies through pinkish to yellowish white. One locality, in a cutover exposing the unit on the northeast limb of the anticline, displays salmon pink and black nodular marble. Dolomite beds up to 20 cm thick occur in lower part of the unit. Angular, centimetre-thick dykes filled by rusty-weathering dolostone cut the ivory marble in some outcrop.

The middle dark-grey marble and the ivory marble are correlated with the Catoche Formation, St. George Group. The ivory marble is stratigraphically and lithologically identical to the Costa Bay member of the formation. The presence of abundant stylolites and its thick, uniform development below the Costa Bay member suggest that the original lithology at the top of the dark-grey marble was a grey peloidal or skeletal grainstone similiar to that described in the Catoche Formation in the Pistolet Bay area throughout the Northern Peninsula (Knight, 1977, 1991; Lane, 1984). The lack of an overlying unit of dolostone suggests that the Aguathuna Formation correlatives were either not deposited or were eroded before the deposition of the overlying carbonate units. Regional stratigraphic evidence such as the thick development of the Aguathuna Formation in the nearby Goose Arm–Old Man Pond area and the presence of dolostone-filled fissures cutting the ivory marble in the map area suggest that there is a marked erosional disconformity at the top of the unit. This sequence boundary is known throughout Newfoundland as the St. George Unconformity (Knight *et al.*, 1991).

Unit 9: Upper Dark-Grey Marble

The uppermost marble unit, which attains about 200 m in thickness, underlies the northeastern-most part of 'Pye's ridge'. Uniform in colour, they consist mostly of dark-grey to grey-cleaved limestone marble having few distinctive features. Cliffs of the marble indicate that bedding is massive and thick but for the most part the marble is intensely veined and sheared especially in its upper part.

Cyclic carbonates mark the basal part of the formation. They outcrop in cutovers just east of the main track across the northeast termination of 'Pye's ridge'. They consist of thick beds of thin stylo-bedded grey limestone marble, overlain by light-grey marble, characterized by white calcite-cemented tubular burrows, in turn overlain by light-grey limestone having white laminar fenestral fabrics. These rocks are also associated with nodular, dolomitic blue-grey marble. Such rocks are characteristic of the Spring's Inlet member of the Table Point Formation (Ross and James, 1987; Knight, 1991) and help to confirm that the upper dark-grey marble is equivalent of the Middle Ordovician Table Point Formation, Table Head Group.

Unit 10: Dark-Grey Phyllite and Marble

At the foot of the northwest slope of the ridge close to 'Helen's brook', a few outcrops of grey phyllite having a well-developed slaty cleavage occupies a structural wedge between the upper dark-grey marble to the south and the faulted boundary with the Old Man Allochthon to the northwest. A single band of recrystallized, dark-grey limestone, dolostone and phyllitic dolostone is intercalated with the phyllites. Each carbonate lithofacies forms units 2 to 5 m thick. The limestone is massive containing folded stylolitic seams. The dolostone is rusty-yellow-weathering, and cut by numerous white quartz veins and quartz–salmon-pink feldspar veins. The phyllitic dolostone is thinly planar stratified and picks up a well-developed cleavage in more phyllitic stratum.

The actual age and correlation of this unit is problematic, but based upon comparisons with rocks outcropping just beneath the Humber Arm Allochthon west of Old Man Pond, it belongs to the Goose Tickle Group. The phyllites were originally shaly flysch of the American Tickle Formation and the carbonates may belong to the Daniel's Harbour member of the same formation (Stenzel *et al.*, 1990).

MELANGE

Unit 11 occurs along 'Helen's brook' in one small area at the north end of the ridge where it overlies structurally, sheared marbles of Unit 9. It consists of polydeformed green-grey and blue-black phyllites with carbonate lenses, disarticulated green-grey, very fine-grained psammite and quartz–carbonate veins.

OLD MAN POND ALLOCHTHON

Unit 12: Old Man Pond Group

Chlorite grade metasedimentary rocks of the Old Man Pond Group outcrop to the northwest of the 'Helen's brook' fault. They include polydeformed pelites and psammites of Cambrian age and older and are not discussed here further.

CARBONIFEROUS

Unit 13: North Brook Formation, Deer Lake Group

Carboniferous sediments occur in two settings: 1) as conglomerate, breccia, sandstone and siltstone overlying unconformably both Paleozoic carbonates and the Old Man Pond Group; and 2) as cave sediments and precipitates principly developed along faults. The presence of these sediments scattered around 'Pye's ridge' indicate that the present topography is essentially an exhumed Carboniferous landscape.

1) Sedimentary Rocks

Remnants of grey, calcareous carbonate breccia and conglomerate occur at several places fringing Pye's ridge. In the south of the ridge flanking Eighth Brook, breccias (Plate 6) comprising locally derived dolomite and lesser marble set in a slightly arenaceous, calcareous matrix lie against an angular Carboniferous paleo-hillside carved out of Unit 3 dolomite. Of special significance in the breccia are the presence of white, fine-grained rhizoliths (Plate 6) (Esteban and Klappa, 1983). The rhizoliths include rhizoconcretions and root petrification and combinations of both. The best examples are stout elongate concretions with root and rootlet tubules well preserved in the circular cross-sections. However, they also occur in the matrix as fine tubules, thin sheets or irregular clots locally attached to the underside of carbonate clasts.

The breccias appear to be a metre or so in thickness and are onlapped by red, friable, small-pebble conglomerate and sandstone that is seen throughout the Eighth Brook area as blocks and weathered rubble as high as 150 m up the present hillside.

Grey calcareous conglomerate also occurs at the bottom of and high on the upper-valley slopes (150 m) of 'Helen's brook'. The conglomerates are several metres thick and consists of locally derived marble clasts 10 to 40 cm in diameter. Almost flat-lying, red-brown, micaceous siltstone and some friable, crossbedded and crudely stratified sandstone occur in pockets along the south slopes of 'Pye's ridge'. These sediments belong to the North Brook Formation of the Deer Lake Group (Hyde, 1979).

Plate 6. *A grey carbonate breccia of Carboniferous age containing angular dolomite marble clasts set in a green-grey calcareous matrix. A long white rhizolith is indicated by an arrow; beside wood road climbing Eighth Brook to the crest of 'Pye's ridge'.*

2) Cave Deposits

Small cave deposits are located along 'Helen's brook' fault and the western, northeast-trending fault that traverses the crest of 'Pye's ridge' (Plates 7, 8 and 9). At 'Helen's brook', several caves (Plate 9) occur at the contact of marble and phyllite and along minor faults that splay off the 'Helen's brook' fault. The caves are up to 30 cm wide and 1.5 m long and are filled by red argillaceous limestone that is either massive or has discordant, stratification. Cave walls are smooth to solution sculptured.

Plate 7. *Botryoidal calcite flowstone with brown, ?dolomitic, banded crust lining the walls of a Carboniferous cave developed along a northeast-trending fault at the crest of 'Pye's ridge'.*

Several cave deposits occur along a 2 km length of the western fault that strikes across 'Pye's ridge'. The caves are narrow, generally less than 30 cm wide and locally can be followed for several metres along the fault zone. Unlike the 'Helen's brook' caves, these caves contain well-developed calcite speleothems that include globular botroyidal forms (Plate 7) having a distinct internal banding, sheeted flowstone that is commonly brecciated and locally incorporated in later cave sediments (Plate 8), and cave pisolites or pearls. In some instances where the cave has dolomite walls, the precipitates are buff-weathered, suggesting a magnesium and possibly ferruginous composition. Cave sediment includes buff- to red-coloured, fine-grained carbonate silt locally enclosing speleothem fragments and carbonate gravel. The latter is locally sheared in one cave (Plate 8). Wall rock is commonly reddened for several metres from the faults and caves.

Plate 8. *Fragments of flowstone (short arrow) incorporated in green to red cave sediment. The Carboniferous cave sediment is faulted (long arrow) and deformed by sinistral shear.*

Plate 9. *Red Carboniferous cave fills (arrows) along small faults and fractures that occur in the footwall carbonates of the 'Helen's Brook' fault (location 11, Figure 4).*

STRUCTURE

The carbonates underlying 'Pye's ridge' are polydeformed (Figure 2). The main structures are the 'Pye's ridge' thrust (PRT), the 'Pye's ridge' anticline (PRA) and the 'Helen's brook' fault (HBF). The PRT outcrops midway along the ridge, repeating the Cambrian part of the succession. The PRA is a northeast-trending and plunging fold that disposes the broad structure of the ridge and folds the PRT. The HBF is a high-angle normal fault that forms the northwest contact of the carbonate sequence with metasedimentary rocks of the

Old Man Pond Allochthon to the northwest. A number of northeast-trending faults, with complex displacement history, traverse the length of the crest of the ridge.

THRUST FAULTS

'Pye's Ridge' Thrust

The PRT carries ribbon limestone and phyllite of Unit 2 over locally recumbently folded and foliated marbles of the interbedded limestone-dolomite marble (Unit 4). At the crest of the PRA, the thrust is gently dipping to the north (attitude based upon foliated footwall marbles just below the thrust plane). The thrust, however, seems to cut up section as it is traced north and south across the ridge. To the north, it swings southwest and steepens to 60°, juxtaposing hanging wall and footwall dolomites (both Unit 3) against one another. Its dip shallows as it swings west to intersect with the northeast-trending HBF. In this westernmost area, the footwall is Unit 2 ribbon limestone and phyllite whereas the hanging wall, although concealed in low boggy ground with no outcrop, is surmized as marbles of Unit 6 and younger, or may be Unit 10 grey phyllite and marble. Unit 3 dolomite forms the hanging wall above footwall rocks of Unit 4 as the PRT is traced southward over the crest of the ridge before it is truncated by a northeast-trending reverse fault.

Thrust zones are characterized by complex structural relationships. The sole of PRT's hanging wall is marked by a well-developed phyllite—limestone marble banding. It is uncertain whether this banding is original bedding or tectonic in origin. This is because in the Goose Arm area such banding is frequently related to transposition of bedding that is almost indistinquishable from the original bedding (Knight and Boyce, 1991). Small-scale folds having a westward and locally eastward sense of vergence and axial planes co-planar with the banding suggest a tectonic origin. Recumbent folds having a west sense of vergence, bedding co-planar S_1 foliation, and a stretching lineation occur in footwall marbles. Locally, the lineation pitches 36° northwest at 320 to 330°.

Evidence of Other Thrusts

Intensely deformed ribbon limestone and phyllite (Unit 2) and pelite—psammite (Unit 1) at the southwest end of 'Pye's ridge' suggest a subcropping thrust below the ridge. In this area, a west-trending, northward-dipping mylonitic foliation occurs in Unit 1 where psammitic folia are deformed as rootless intrafolia folds (Plate 1). Folded psammite and pelite in low-strain zones in the mylonite indicates west-verging compressive stress. Unit 1 also cores small, outcrop-sized, boomerang-shaped interference fold patterns within Unit 2 west of Eighth Brook. Complex folding of banded phyllite and ribbon limestone occurs in the same area. Small-scale, tight, isoclinal, recumbent folds folded by D_2 folds are common. The banding folded by these isoclinal structures, however, may not be bedding but transposition bedding, suggesting that at least multiple phases of deformation produced the D_1 fabric. The presence of associated sheath folds supports this hypothesis. Decametre-scale recumbent folds are also common (Plate 10).

Plate 10. *Polydeformed limestone and phyllite of Unit 2 at the southwest end of 'Pye's ridge'. A recumbent fold closure containing the j-shaped eye of an earlier fold occurs to the left of the hammer. The hammer is parallel to a foliation that is axial planar to the second-phase recumbent fold. S_2 and F_2 folds are deformed by a F_3 fold associated with a cleavage (arrow) that dips into the hill and to the right and has a right-handed sense of shear.*

A third thrust is likely at the northeast end of the ridge in the vicinity of North Brook. Dark-grey marbles of the upper part of Unit 9 in this area carry a prominent, generally shallowly dipping shear fabric, which is folded by D_2 folds and offset by D_3 fractures. Reconnaisance of this area and the ridge to the northeast of 'Pye's ridge' suggests that stratigraphy, including at least the ivory marble (Unit 8) and above, is repeated in the hanging wall of the thrust. Structural wedges of Unit 11 melange structurally overlie the marble in the footwall of the postulated thrust.

Although the sense of vergence and direction of lineations suggest west- to northwest-directed thrusting, the complexity of structure in the thrust zones makes this conclusion tentative. However, it is supported by a small-scale horse structure (Plate 11) recognized in thin-bedded marble of Unit 4 in the hanging wall of the PRT that clearly indicates a northwestward sense of shortening.

FOLDS

The 'Pye's ridge' anticline is one of two folds that deform the sequence as a whole. Both are northeast-trending and plunging folds associated with moderately northwest-dipping cleavage. Locally, the southeast limb of the PRA is overturned. A related syncline occurs in the northeast corner of the map area. It is mostly truncated and obscured by Carboniferous and recent sediment cover as it is traced southwest along the southern side of the ridge.

FAULTS

Two northeast-trending faults that dip at 55 to 71° northwest, traverse the crestal area of the anticline. The

Plate 11. *Small-scale horse structure developed in sheared off-white marble of Unit 4. The sense of thrusting is to the northwest (left of photograph); near location 3 of Figure 4.*

western fault is characterized by a hanging wall shear zone several metres wide (Plate 12). Footwall dolomites are strongly shattered. To the southwest, the shear zone passes into a brecciated zone that is overprinted by red colouration and postdated by Carboniferous cave deposits that are locally deformed by sinistral displacement (Plate 8). Drag folds on this fault indicate reverse motion but present disposition of units across the fault is more compatable with normal displacement. This is reflected in the numerous small-scale indicators such as small folds, C−S-like fabrics and rotated boudins indicate that the main sense of movement across the shear zone is dextral, or down dip. Locally, some small folds and C−S fabrics also give a sinistral sense of shear. The order of movement is not presently understood. Several sets of dilation fractures occluded by calcite cut the shear foliation. Some of these structures give a sinistral sense of movement. The eastern fault has a narrow 40-cm shear zone and dextral sense of shear.

'Helen's Brook' Fault Zone

The 'Helen's brook' fault strikes 225° and dips 70° northwest (Plate 13). It seperates Old Man Pond Group from sheared upper dark-gray marble and phyllite of Unit 10. The psammites and pelites of the Old Man Pond Group have an early schistosity that is axial planar to recumbent folds. They are also cut by a later crenulation cleavage and are folded about north- to northwest-trending axes. Close to the fault, S_1 schistosity is cut by fault-related fractures striking 210° and dipping 35° west. They displace the schistosity and indicate a normal sense of shear.

Footwall rocks also suffered several deformations. Marbles have a steeply dipping foliation that locally, in the southern part of the fault zone, strikes 214 to 245° and dips west at 31 to 52°. The variation in strike is attributed to faults that splay off the HBF at 250 to 255° strike and dip 54 to 60° northwest. They repeat the contact of the foliated marble and the overlying Unit 10 phyllite with marble. Sense of displacement on these splay faults is reverse. Fabrics in the foliated marble and the overlying phyllites are discordant to each other by 90°, suggesting that their pre-faulting relationship is structural not stratigraphic. The marble foliation is associated with a steeply dipping stretching lineation that plunges at 51° and trends 326°. Late fractures cutting the phyllite and marble trend 230°, dip west at 51° and have a sense of shear compatable with normal displacement on the HBF. Similiar structural relationships occur in a number of outcrops of upper dark-grey marble and Unit 10 phyllites in the north of the fault zone.

Plate 12. *Mylonitic shear zone developed in the hanging wall of a northeast-trending fault cutting the crest of the 'Pye's ridge' anticline. The fault plane is in the lower part of the plate. En-echelon and irregular calcite veins cut the foliation; Near location 13, Figure 4.*

Plate 13. *Steeply northwest-dipping 'Helen's brook' fault (left of hammer), 'Helen's brook'. Psammites and pelites of the Old Man Group have a flat-lying schistosity deformed by normal-sense drag folds and C−S fabrics. The footwall consists of Unit 10 phyllites.*

OTHER STRUCTURES

Kink bands trending 232° and dipping at 42° northwest are rare in the marble of Unit 4. Numerous small calcite- and quartz-filled dilation fractures occur throughout the succession but are not described here.

INTERPRETATION OF THE STRUCTURAL HISTORY OF 'PYE'S RIDGE'

Three main phases of deformation effect the sequence. The first is related to thrusting with zones of high-ductile strain, developed particularly in Units 1 and 2. Complexity of structures and fabrics in these zones indicate that they hide multiple deformation themselves. Stratigraphic relationships associated with the PRT suggest that it structurally transgresses stratigraphy and was probably a ramp.

Although only the PRT is mapped, a lower thrust is postulated to occur below the ridge. In addition, reconnaisance of the northeastern end of 'Pye's ridge' and the adjoining ridge to the northeast suggests that there is another thrust between the ridges. This thrust duplicates the upper part of the carbonate sequence and overlying flysch by overthrusting the Ordovician rocks and melange of the northeast end of 'Pye's ridge'. A widely developed early foliation and mylonitic fabric deformed by later structures in the top of Unit 9, and in Unit 10 along the northwest side and northern end of the ridge, indicate that a thrust or thrusts occur above the upper thrust slice of the ridge. In one case, melange structurally overlies sheared, upper dark-grey marble in the north of the ridge suggesting that this relationship may in part be related to emplacement of the allochthons above the platform sequence. Discordant structural relationships between phyllite (originally flysch) of Unit 10 and Unit 9 marbles also indicate décollement between the platform and overlying flysch.

These structural relationships imply an imbricated thrust stack assembled during or after the emplacement of the allochthons. The structural fabrics of the polydeformed metasedimentary rocks of the lower thrust slice indicate polydeformation during thrusting but whether this relates to Taconian allochthon emplacement or to later Acadian shortening of an already assembled package is yet to be resolved. If the allochthon is the hanging-wall plate, the underlying disrupted autochthonous sequence may be imbricated as part of a duplex structure. In this situation the 'Pye's ridge' anticline could have formed as a hanging-wall anticline to a ramp below the lower thrust.

However, southeast-verging, northeast-trending folding and cleavage is thought to be a second phase of deformation. This is suggested by the folding of the PRT and by the presence of several scales of interference fold patterns that appear to postdate the thrustal shortening. Northeast-trending high-angle faults having evidence of reverse displacement are probably related to this event.

The third deformation is associated with the vertical juxtaposition of the Old Man Pond Allochthon against the carbonate sequence across the HBF. It is characterized by deformation of earlier fabrics by downdip extension along fractures parallel to the fault in both hanging wall and footwall rocks. Steepening of dips in the northwest limb of the 'Pye's ridge' anticline may be related to this. Similiar down-dip extensional fabrics associated with northeast-trending faults that traverse the crest of the 'Pye's ridge' anticline are probably related to reactivation of these D_2 faults during this phase.

Minor deformation of Carboniferous cave deposits along the northeast-trending faults indicates minor brittle deformation of the sequence in Late or post-Carboniferous times. Rare kink bands may be related to these events

The three main deformations are interpreted to reflect: 1) westward-verging shortening during or following the westward emplacement of the Old Man Pond Allochthon; 2) southeast-verging backthrusting; and 3) normal sense juxtaposition of the Old Man Pond Allochthon adjacent to the Paleozoic marbles. These three phases all predate the Carboniferous.

The three deformations are comparable to the structural chronology described in the Old Man Pond area by Waldron and Milne (1991). They suggested that the structural relationships at the west end of the Old Man Pond Allochthon indicate the emplacement of Reluctant Head Formation and overlying carbonate platform above the allochthon. From the work at 'Pye's ridge', it is clear that the carbonate facies are very similiar to those described at Old Man Pond west (Knight and Boyce, 1991) and that the structural relationships in the area of the ridge show the allochthon was downdropped to occupy its present position. Knight and Boyce (1991), and reconnaisence mapping by Knight in 1991 in the Old Man Pond west area, also suggest that the Reluctant Head thrust slice is emplaced upon lower thrust slices of platform carbonates. The thrusts first verged to the west, and then were folded by east-verging structures. This is similiar to the chronology described for D_1 and D_2 structures at 'Pye's ridge'. In addition, to the south of the western end of Old Man Pond, the carbonates of the Table Head Group and overlying Goose Tickle flysch lie structurally beneath the allochthon. This indicates that the structural relationships at the west end of Old Man Pond are the result of east-verging back thrusting as suggested by Knight and Boyce (1991).

MARBLE DEPOSITS

The metacarbonates that underlie 'Pye's ridge' have a number of prospective deposits of dimension stone, some of which are currently being evaluated for commercial exploitation. Two stratigraphic intervals that provide very contrasting styles of marble are presently the main exploration targets. Unit 4 is dominated by several white varieties of marble that include a number of bodies of arabescato (Plate 3). The stratigraphic interval, which always immediately succeeds Unit 3 dolomite, is repeated by the PRT and potential dimension-stone deposits are present in both thrust slices. The other stratigraphic target is marked by grey marbles having a distinctive stylolitic pattern (Plate 5). They occur in the upper part of Unit 7 and underlie a large area along the southeastern slope of the ridge. Other attractive marbles occur in Units 2 (Plate 2), 8 and 9 and these are discussed along with marbles that have been effected by faulting and Carboniferous diagenesis to give distinctive qualities.

In general, the marbles of 'Pye's ridge' are fine to medium grained, having good crystallinity and are compact and massive. Natural structures to exploit the marbles include bedding and cleavage planes. Comparison of the 'Pye's ridge' marbles with those of Italy is based upon a recent visit to the quarries of Carrarra, the marbles exhibited at the Museo del Marmo at Carrarra and the Italian Marble–Technical Guide (Italian Institute for Foreign Trade, 1982).

WHITE MARBLE OF UNIT 4

White, ivory, pale blue-grey, light-grey and locally pink marbles occur in the unit (locations 1 to 4, Figure 4). The white varieties *(bianco)* are rarely truly brilliant white but several outcrops indicate a number of varieties are likely from white alone to white having a faint grey veining *(bianco venato)* to dappled off-white and grey, which appears to reflect the metamorphism of microbial mounds. Strongly foliated white marble (Plate 14) usually has streaky colour variation that can include folds and resembles the variety *nuvolato.* Most spectacular amongst these marbles are several bodies of *arabescato* (Plate 3). This is a marmorized limestone conglomerate and breccia, several metres thick, that is fairly consistently mapped within a few metres of the base of Unit 4. White marble and some dolostone clasts are set in a dolomitic to calcareous matrix that has sericitic seams. Inhomogenous strain throughout the area means bodies range from almost undeformed conglomerate to conglomerates having strongly flattened clasts. The latter are common in the upper thrust slice (location 3) and also occur in the lower thrust slice where the conglomerate is recumbently folded below the PRT (location 4).

Other varieties in Unit 4 that appear attractive are the off-white marbles that retain crossbeds and a number of beds of a fine-grained, smooth pale blue-grey marble. At the base of the Unit 4, just below the *arabescato* interval, pure ivory to cream marble pass laterally into creamy dolomite marble. The latter has an uniformly fine-grain size and is very massive and hard.

GREY MARBLE OF UNIT 7

Grey marble (location 5) commonly refered to in Italian quarries as *bardiglio* is both thick (many tens of metres) and extensive along the southern slope of 'Pye's ridge'. The variety is characterized by the presence of a dark-grey to black stylolite (Plate 5) that essentially picks out bedding in the rock. The stylolites are folded at different scales and this combined with the original (pre-deformation) intersecting nature of these solution structures makes this a very attractive rock, not significantly different from *bardiglio imperial* of the Massa area, Italy. The outcrops in the area indicate metre-thick bedding and good consistency of material.

MARBLE OF UNIT 2

Marble in the upper part of Unit 2 (locations 6 and 7) is banded. They consist of blue-grey calcareous bands having thin phyllitic partings and seams. The banding varies from planar to folded with small offsets as well as fine calcite veinlets common. The unit is well developed extensively in both lower and main thrust slice and the quarryable interval is likely to exceed tens of metres.

IVORY MARBLE OF UNIT 8

This 20-m-thick stratigraphic unit is consistently developed around the PRA in the upper part of the main thrust slice. The gently to moderately dipping unit is characterized by ivory to white colouration and green, sericitic to buff, dolomitic stylolite seams parallel to bedding. Based on comparison with the equivalent unit elsewhere in western Newfoundland, the unit is likely to be quite pure especially in the middle and upper parts. A conjugate set of joints is developed in the unit around the PRA. The most interesting marble in this unit occurs in the vicinity of the synclinal closure (location 8). There the marbles are off-white, white and light-grey with a very attractive dark-grey to black stylolite, not unlike the Italian variety *uliano venato* and *calda 'D'.*

MARBLE OF UNIT 9

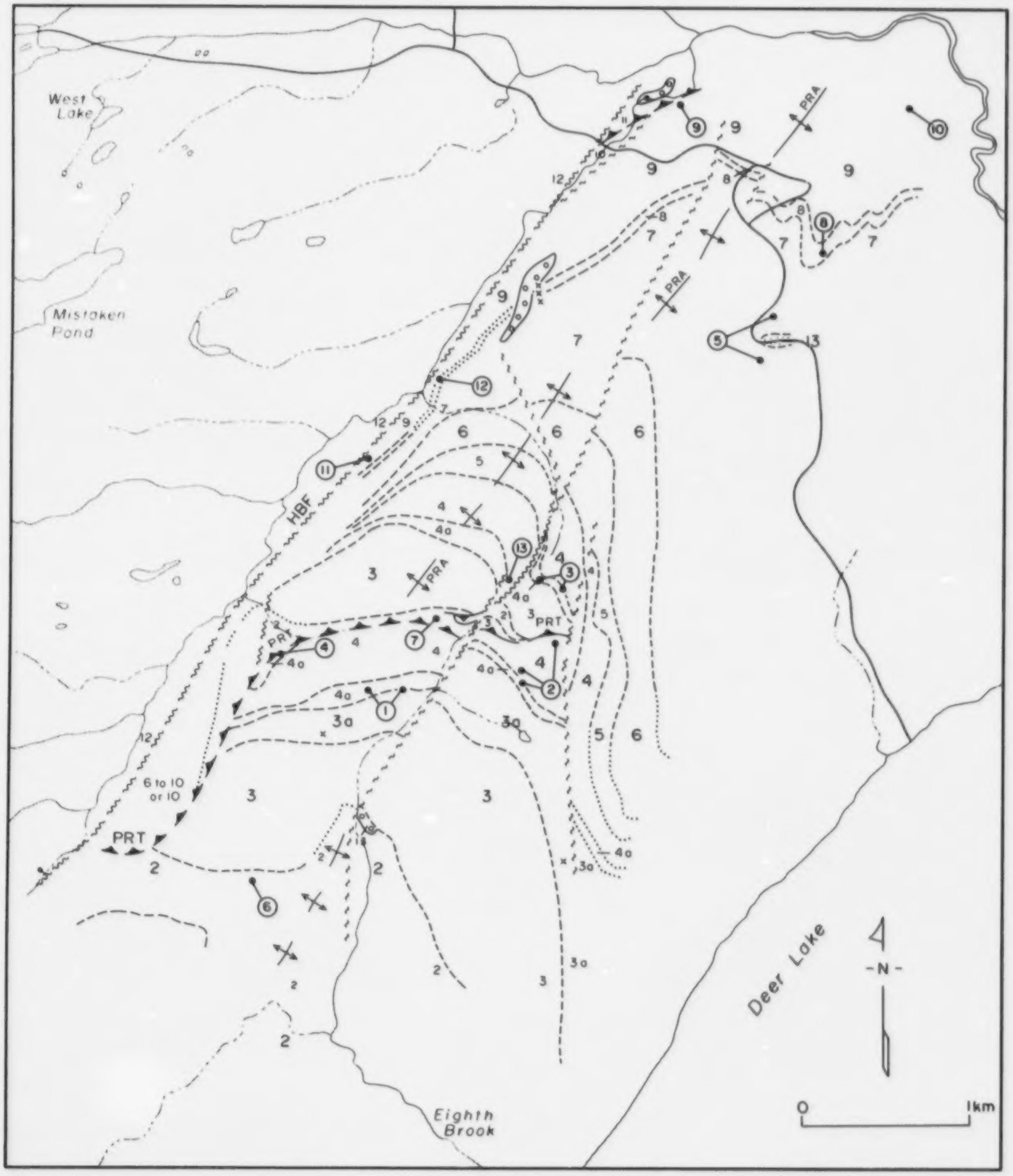
Although Unit 9 comprises a dark-grey marble of probable little decorative interest, modification of the unit by intense deformation has added some attractive variants. At locations 9 and 10 in the northeast of the area, several phases of shear superimposed upon black to buff, argillaceous and dolomitic stylo-bedding and burrow fabrics have created an impressive *bardiglio scuro* (Plate 15). To the imaginative eye, the textures are reminisant of dense leafy foliage blown by a gale. Retention of light-grey to white burrow and fenestral fabrics in grey marbles in the basal cyclic part of Unit 9 are similiar to *bardiglio nuvolato* from Carrarra, Italy and will require additional assessment.

OTHER MARBLE

Marbles preserved close to the HBF along the northwestern side of 'Pye's ridge' are commonly characterized by a strong foliation that verges on a mylonitic fabric. The resulting varieties of marble vary in colour from off-white to dark-grey but all have the resemblence of the Italian variety *striato* (locations 11 and 12). Strongly sheared grey marble cut by an abundance of white calcite veins is common along stretches of the hanging wall associated with the northeast-trending fault that strikes along the crest of 'Pye's ridge'.

PINK MARBLES, VEIN AND BRECCIA MARBLES ASSOCIATED WITH CARBONIFEROUS OVERPRINT OF FAULTS

These marbles are best developed close to the western northeast-trending fault that intersects the marbles at the crest of 'Pye's ridge' (location 13). Hanging-wall rocks in the area are commonly red to pink in colour and vary from intensely foliated and veined varieties to those that are brecciated. The


West Lake
Mistaken Pond
Deer Lake
Eighth Brook
HBF
PRA
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①
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④
⑤
⑥
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⑧
⑨
⑩
⑪
⑫
⑬
2
3
3a
4
4a
5
6
7
8
9
12
13
6 to 10 or 10

Figure 4. *Location map of marbles described in the text. • location of marble deposit.*

LEGEND (for Figure 4)

CARBONIFEROUS

13 *Grey conglomerate–breccia, red conglomerate, sandstone and siltstone*

?PRECAMBRIAN–CAMBRIAN

OLD MAN POND ALLOCHTHON

12 *Psammite and pelite*

?ORDOVICIAN

11 Melange

MIDDLE ORDOVICIAN

GOOSE TICKLE GROUP

10 *Dark-grey phyllite with marble; grey phyllite, thin-bedded and conglomeratic limestone and dolostone*

TABLE HEAD GROUP

9 UPPER DARK GREY MARBLE *Thick-bedded, massive dark-grey to grey limestone; cyclic with fenestral fabrics and rare dolostone at the base*

LOWER ORDOVICIAN

ST. GEORGE GROUP

8 IVORY MARBLE *Ivory- to white-coloured, thinly bedded marbles having green, dark-grey and dolomitic, stylolitic partings; minor dolostone beds at the base*

7 MIDDLE DARK GREY MARBLE *Burrow-dolomite-mottled, thick bedded and stylolitic*

6 VARICOLOURED MARBLE *Vari-grey to white, burrow-mottled, mounds, dololaminite*

5 LOWER DARK-GREY MARBLE *Thick-bedded, burrow-mottled, mounds*

CAMBRIAN

PORT AU PORT GROUP

4 LOWER LIMESTONE–DOLOMITE MARBLE *White, blue to light-grey marble, dololaminite, mounds, crossbedded;* **4a**, *conglomeratic*

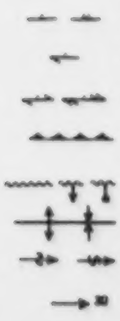
3 DOLOMITE MARBLE *Light-grey to cream, massive, thick bedded;* **3a**, *cherty in upper part*

2 RIBBON LIMESTONE–PHYLLITE *Blue-grey, thin-bedded marble having phyllite and dolomitic beds and partings; some grainy and conglomeratic beds*

LABRADOR GROUP?

1 PELITE AND PSAMMITE *Green-grey to grey pelite having white psammite laminae*

SYMBOLS



Bedding, tops known, unknown

Co-planar bedding-cleavage; schistosity

Cleavage, D_1, D_2

Thrust fault

High-angle fault, dip, downthrow

Fold axes, anticline, syncline

Minor folds, plunge, upright to inclined axis, recumbent to reclined

Lineation, plunge

Plate 14. *White marble showing a crenulated, yellow stylolitic fabric cut by a transposition fabric axial planar to the crenulation. Pure marble is developed in the zones of transposition. Upper thrust slice, 'Pye's ridge' near location 3, Figure 4.*

Plate 15. *A dark-grey marble (Unit 9) with well-developed shear fabric gently crenulated by obliquely dipping fractures (arrows) with a normal sense (left-handed) of shear; 'Helen's brook', north end of 'Pye's ridge'.*

latter are characterized by foliated, often off-white marble fragments cemented by a red to yellow calcite. The colouration and possibly the brecciation is Carboniferous in age. The zone attains several metres in thickness.

MARBLE DIMENSION-STONE PROSPECTS ELSEWHERE IN WESTERN NEWFOUNDLAND

LOWER PALEOZOIC MARBLES

Marbles in the Deer Lake area form part of a narrow belt that extends from Canada Bay in the north to Humber gorge near Corner Brook. This belt occupies the most easterely edge of the lower Paleozoic platform sequence. The marble belt abuts structurally against psammite and pelitic rocks of the inner domain of the Humber Tectonostratigraphic Zone of Williams (1978) in the south and the Hare Bay Allochthon in the north.

Several marble deposits occur in the Canada Bay—Coles Pond area on the Northern Peninsula (Howse, 1986, 1989) where they have been assessed for both dimension stone and industrial filler. Unlike the 'Pye's ridge' marble, those of Canada Bay are finer grained and commonly display buff to green phyllitic seams parallel to bedding. The ivory to white marbles are believed to occupy the same stratigraphic interval as those in the Deer Lake area. Zones of strongly foliated marble are also common giving a good '*striato*'-like stone. Southwest of Deer Lake, the belt should continue southwestward from the Humber gorge adjacent to the Corner Brook Lake Thrust. The stratigraphy of this area is similiar to that on 'Pye's ridge' suggesting good ground for prospecting.

CARBONIFEROUS DEPOSITS

The presence of Carboniferous cave fills along faults is likely in the southern part of the marble belt. That these deposits can be spectacular is shown by the deposit recently destroyed during construction of the Trans-Canada Highway through the Humber gorge. The deposit, which occurred along the south side and the west-central part of the gorge trended parallel to the west-flowing river, and is hosted by massive Paleozoic metacarbonate that structurally lies just east of banded phyllitic marbles of the Reluctant Head Formation. The cave fill was several metres wide and thick and was traced for tens of metres. It was characterized by a dyke of breccia (the cave fill) flanked by a zone of extensively fractured wall rock (cave wall). Breccia clasts and wall rocks are both reddened and the clasts and fractures surrounded and occluded respectively by bright white calcite cements that commonly display faint red colour banding. Well-cemented red sediment infilling open spaces also adds to the attractiveness of the stone in some cave fills of the Deer Lake area. Dimension stone quarried from similiar cave deposits figured prominently at the 1991 Marmomachine Trade Fair for dimension stone at Verona, Italy, suggesting that these types of deposits should not be overlooked and dismissed casually in the future.

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INTERNAL AND EXTERNAL RELATIONSHIPS OF THE SOUTH LAKE IGNEOUS COMPLEX, NORTH-CENTRAL NEWFOUNDLAND (NTS 2E/5, 6): ORDOVICIAN AND LATER TECTONISM IN THE EXPLOITS SUBZONE?

B.H. O'Brien
Newfoundland Mapping Section

ABSTRACT

The South Lake Igneous Complex of north-central Newfoundland occurs in a portion of the Exploits Subzone (Dunnage Zone) that is characterized by Early and Middle Ordovician plutonic and volcanic rocks, which formed within the Iapetan oceanic crust, and by Late Ordovician sedimentary rocks that accumulated above this substrate. Bodies of gabbro, diorite and tonalite, along with several swarms of mafic dykes, constitute the igneous complex, and make up distinctive tracts of ophiolitic and magmatic arc rocks. Their interrelationships establish, and relative ages bracket, several Ordovician periods of ductile deformation, dynamothermal metamorphism and mesozonal plutonism that were internal to the South Lake Igneous Complex. They also point to the important role of shear zones in ophiolitic basement and their control on the tectonic development of younger rocks.

The external relationships of the Lower–Middle Ordovician South Lake Igneous Complex with the Middle Ordovician Wild Bight Group and Upper Ordovician Point Leamington greywacke are interpreted to mean that the complex was exhumed and eroded by the Ashgill, having been accreted to Exploits Subzone rocks of overlapping age by the Llanvirn. Structurally, the South Lake Igneous Complex is a parautochthonous unit, externally bounded by high-angle reverse faults of possible Late Silurian age.

OBJECTIVE

The northern portion of the South Lake Igneous Complex and adjacent parts of the Point Leamington greywacke and Wild Bight Group were examined as part of a regional mapping project in central Notre Dame Bay (Figure 1). The data obtained from these rocks will be integrated with those in other regions of the Exploits Subzone in a regional study that, amongst other objectives, will deal with the interrelationships of rocks of Early, Middle and Late Ordovician age (O'Brien, 1991a,b; Dec *et al.*, *this volume;* Williams *et al.*, *this volume*).

RELATIONSHIPS INTERNAL TO THE COMPLEX

Between Martin Lake and Cramp Crazy Lake, the South Lake Igneous Complex (Dean, 1977) is composed of several types of plutonic rocks and a variety of minor mafic intrusions, all of which are variably deformed and metamorphosed. Mappable bodies of mafic to felsic plutonic rocks (Figure 2) are, from oldest to youngest, metagabbro, hornblende diorite, quartz tonalite and quartz diorite. Lorenz and Fountain (1982) considered these rocks to be representative of the entire complex.

THE OPHIOLITE TRACT

The oldest exposed part of the South Lake Igneous Complex, originally referred to as the South Lake Ophiolite (Dean, 1977), is restricted to a relatively small tract in the northern part of the complex.

Metagabbro (Unit 1)

Identified as Unit 1 metagabbro in Figure 2, the ophiolitic tract consists of northeast- to east-trending zones of cumulate-layered, schistose and flaser-banded gabbro. That intrusion of younger phases of this purported ophiolite (Dean, 1977) accompanied deformation and metamorphism of its older phases is witnessed by the field relationships of *lit-par-lit* leucogabbro veins to variably mylonitized metagabbro (Plate 1). Sheeted mafic dykes, which constitute the oldest recognizable swarm in the South Lake Igneous Complex (Table 1), were emplaced into wrench-type shear zones,

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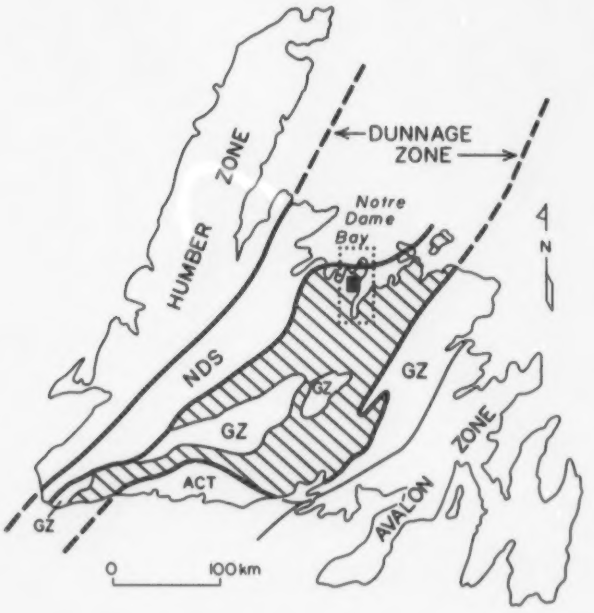


Figure 1. *Location of the study area illustrated in Figure 2 (solid square) within the context of the central Notre Dame Bay mapping project (dotted square). The Exploits Subzone of the Dunnage Zone is indicated by dashed lines. The abbreviations NDS, GZ and ACT represent the Notre Dame Subzone, the Gander Zone and the Avalon Composite Terrane, respectively.*

synkinematically, along with late magmatic veinlets (Plate 2). Such variably sheared dykes, although discordant to the regional structural grain of the ophiolitic metagabbro, were preferentially intruded into subvertical, northwest-oriented, ductile shear zones that developed contemporaneously with dyke-swarm emplacement (Plate 3). In this regard, Unit 1 of the South Lake Igneous Complex is typical of certain zones in the lower parts of fossilized sections of oceanic crust (Dunning, 1987), and structurally similar to preserved fragments of upper mantle that evolved near paleo-transform faults (Suhr *et al.*, 1991).

Engulfment of small fragments of Dunnage Zone ophiolites by syntectonic, dioritic to tonalitic batholiths is a typical feature of the Taconian magmatic arc in southwestern Newfoundland (Dunning and Chorlton, 1985). In the South Lake Igneous Complex, subvertical northeast-trending mylonite zones, each several metres wide, rework metagabbro and originally discordant mafic dykes, where such ophiolitic rocks locally form the complex's western margin (Figure 2). Mylonite zones of this age are no younger than the faulted contact with external units and, based on regionally crosscutting intrusive contacts, are apparently older than some of the other plutonic rocks that made up the complex.

THE MAGMATIC ARC TRACT

Volumetrically, most of the northern part of the South Lake Igneous Complex is made up of diorite and tonalite intrusions. Hornblende diorite, blue-quartz tonalite and quartz diorite are herein distinguished as Units 2, 3 and 4, respectively (Figure 2).

Hornblende Diorite (Unit 2)

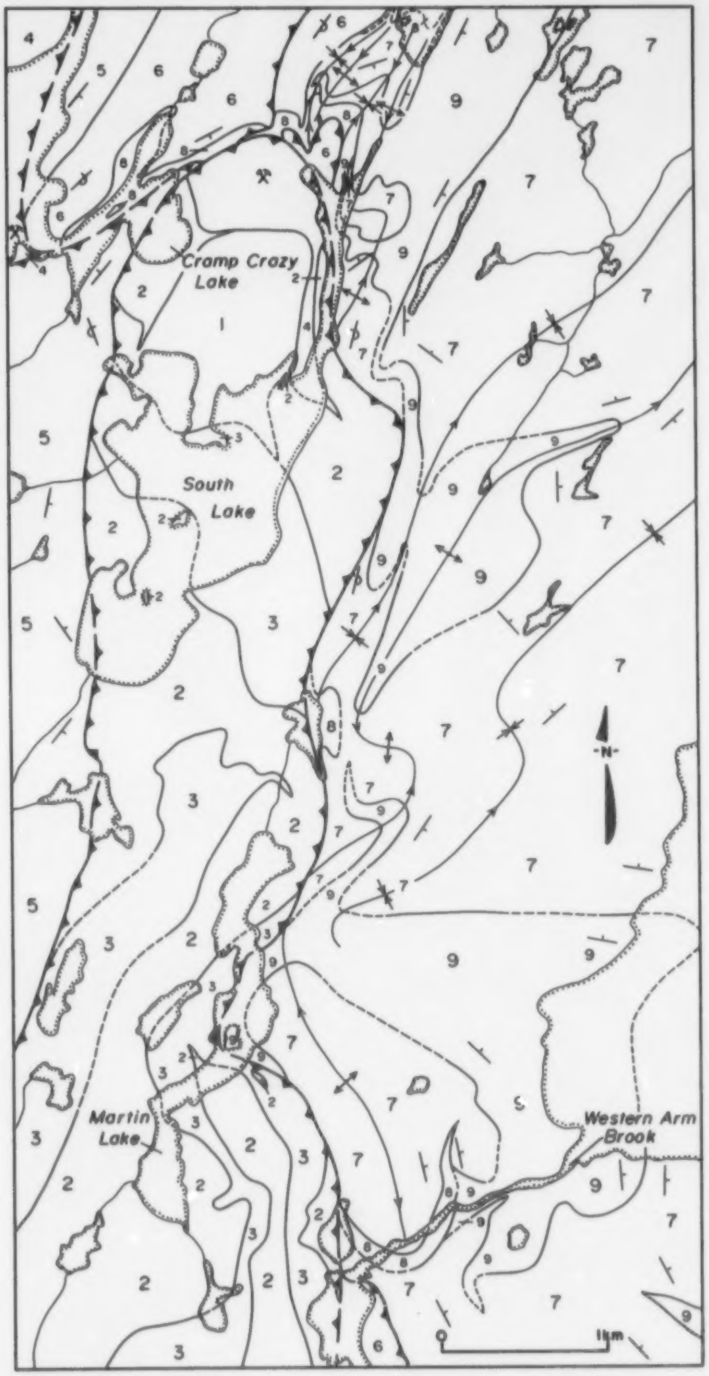
The oldest unit of the purported arc-related suite is hornblende diorite, which crops out as northeast- and northwest-trending, subvertically dipping bodies throughout the South Lake Igneous Complex. Because Unit 2 is seen to occur as septa and as screens, it possibly may have once been a more contiguous entity. Hornblende diorite varies from fine to coarse grained, equigranular to schlieren-textured, and commonly contains mafic pegmatites in diffuse pods or cavities. Most bodies of Unit 2 are unfoliated or weakly schistose, and are rarely altered. Hornblende diorite is host to variably sheared, mafic–felsic, composite dykes (Table 1), which are localized in small northwest-trending trains near segments of the intrusions rich in accidental and cognate xenoliths. Such assemblages of dykes and xenoliths are most common in the northern part of the complex; however, they are not present in all of the hornblende diorite bodies in the area.

A narrow unfoliated appendage of the northwesternmost body of hornblende diorite was apparently intruded into one of the northeast-trending shear zones that rework dyke swarm 1 (Table 1; Figure 2). The main part of this hornblende-diorite body, which crosscuts the complex's oldest dyke swarm as well as significant expanses of flaser gabbro, contains large xenoliths of previously altered and strongly foliated material of felsic composition. The included tectonites are similar to those found in adjacent but younger parts of the complex, suggesting persistent sampling of the same source rock at depth.

Hornblende diorite is closely associated with quartz tonalite throughout that portion of the South Lake Igneous Complex shown in Figure 2. Mafic–felsic composite dyke injection in hornblende diorite might herald blue-quartz tonalite intrusion and indicate that diorite and tonalite are largely comagmatic in origin.

Quartz Tonalite (Unit 3)

The most lithologically distinctive and physiographically prominent unit of the South Lake Igneous Complex is blue-quartz tonalite. Unit 3 tonalite is intruded as a series of anastomosing and tapering sheets in metagabbro and hornblende diorite (Figure 2). Where the complex is composed of alternating diorite septa and tonalite screens, the latter typically contain elongate, variably sized, hornblende-diorite fragments, and the former are commonly concordantly injected by tonalite stringers and overgrown by metasomatic feldspars. Near Martin Lake, ductile shear zones in diorite septa are absent in adjacent tonalite screens, possibly implying that the initiation of diorite deformation may have preceded tonalite intrusion in some areas.



LEGEND

UPPER ORDOVICIAN

POINT LEAMINGTON GREYWACKE

9 *Pebbly wacke–polymictic conglomerate lenticles (pre–D. complanatus Zone ?)*

8 *Grey shale-dominant lenticles (D. clingani Zone ?)*

7 *Unseparated turbidite sandstone units (D. clingani Zone to D. anceps Zone ?)*

MIDDLE ORDOVICIAN

WILD BIGHT GROUP

6 *Red, green and grey, interbedded chert and argillite*

5 *Tuffaceous epiclastic wacke; minor tuff containing mafic and felsic volcanic pyroclasts*

LOWER–MIDDLE ORDOVICIAN

SOUTH LAKE IGNEOUS COMPLEX

4 *Quartz diorite*

3 *Quartz tonalite*

2 *Hornblende diorite*

1 *Metagabbro*

KEY

Geological boundary (defined, approximate)........

Bedding (inclined, overturned)............................

Anticline..................................

Syncline..................................

Fold axial trace (plunge direction indicated).............

Thrust fault (barbs on upthrown side).....................

Base-metal showing................ ⚒

Figure 2. *Geological map of a portion of the Lower–Middle Ordovician South Lake Igneous Complex and adjacent parts of the Middle Ordovician Wild Bight Group and Upper Ordovician Point Leamington greywacke.*

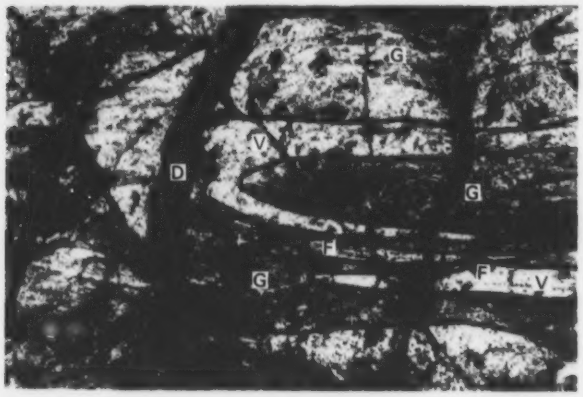


Plate 1. *Host-rock foliation (marked by symbol) is axial planar to isoclinal folds of* lit-par-lit *leucogabbro veins (V) intruding metagabbro (G), indicating a close relationship between magmatism and deformation in this presumed ophiolitic part of the South Lake Igneous Complex. Note that the leucocratic veins infill what were originally a series of branched, asymmetrically forked, shear fractures (F). A mafic dyke (D), belonging to the oldest swarm in the igneous complex, is relatively late in this sequence of events. The dyke illustrated is typical of most of oldest swarm in that it trends across the structural grain of ophiolitic metagabbro but, just outside of the field of view, locally forms T-shaped abuttments with concordant members of this suite of mafic intrusions.*

Generally, quartz tonalite is well foliated but poorly lineated; the fabric defined by a shape fabric in blue-quartz ribbons and by a mineral fabric in aligned ferromagnesian minerals. In both deformed and undeformed states, tonalite bodies are host to the second major swarm of mafic dykes in the South Lake Igneous Complex. However, the initial and waning phases of this dyking event may affect older and younger members of the complex (Table 1).

Undeformed, coarse-grained, quartz-porphyritic tonalite has irregular clots of much finer, ophitic intergrowths of hornblende and plagioclase. This early-crystallized mafic material comprises discrete trains of cognate xenoliths, which are very common in all of the tonalite bodies. They are associated with swarms of distorted, distended and disaggregated (backveined) mafic dykes, especially well developed along intrusive margins of northwest-trending segments of tonalite bodies (Figure 2). The deformation of cognate xenoliths and mafic dykes within tonalite overlapped the deformation of the material backveined from these intrusions. However, there are also numerous localities where strong quartz-ribbon shape fabrics in tonalite are truncated by undeformed mafic dykes.

Accidental xenoliths of mafic, silicic and quartzofeldspathic rocks are found in blue-quartz tonalite. Some of the quartzofeldspathic xenoliths display complex fold-interference patterns and apparently reflect entrapped paragneiss (Plate 4). Though relatively rare, xenoliths of paragneiss may suggest that ophiolitic rocks were overlain by or tectonically intercalated with siliciclastic or epiclastic rocks prior to their intrusion by quartz tonalite. How early (pre-hornblende-diorite) phases of mylonitization might relate to such events, if at all, is unknown. However, xenoliths in which a mafic dyke crosscuts flaser gabbro are present within undeformed blue-quartz tonalite. This indicates that Unit 3 is not allochthonous with respect to the stitched Unit 1–Unit 2 assemblage, and that displacement of ophiolitic metagabbro during the ascent of these arc-related plutonic rocks was regionally insignificant.

Quartz Diorite (Unit 4)

The northernmost part of the South Lake Igneous Complex is underlain by quartz diorite (Unit 4), the youngest member of the complex. For the most part, this medium-grained equigranular rock is rarely foliated and contains uncommon mafic dykes. Internally, the main intrusive bodies are locally saussuritized, sericitized and pyritized. Also, small apophyses of quartz diorite are apparently associated with sulphide-rich alteration zones in those members of the complex that host them.

The margin of the quartz-diorite intrusion north of South Lake is mapped, in part, as transecting the intrusive contact between hornblende diorite and ophiolitic metagabbro and, in other localities, as occupying that boundary (Figure 2). Where quartz diorite net-veins hornblende diorite near their mutual northwest-trending contact, the younger body is observed to have xenoliths of undeformed portions of the older body. Here, medium-grained quartz diorite passes gradationally into fine-grained, porphyritic, marginal phases of this intrusion.

On the shore of South Lake, a north-trending sheet of unfoliated quartz diorite, which crosscuts sheared, northwest-trending Unit 1 dykes, contains accidental xenoliths of mylonite (Plate 5). Distinctive epidote-rich bands within the mylonite suggest derivation, at least in part, from rocks of mafic composition. Near Cramp Crazy Lake, unaltered and unfoliated quartz diorite has xenoliths of highly sheared and partly silicified, blue quartz-bearing, felsic rocks. This material may represent either mylonitic granitoids or strongly deformed, extrusive rocks of felsic composition. The unaltered protolith of these latter xenoliths might be found in adjacent rock groups (Unit 5 of the Wild Bight Group), in older parts of the South Lake Igneous Complex, or in a yet older exotic suite of entirely unrelated rocks.

THE COMPLEX IN SUMMARY

Episodes of persistently ductile and transiently brittle deformation and associated periods of dynamothermal metamorphism were coeval with much of the plutonism in the South Lake Igneous Complex. A long-lived, crustal-scale, vertical shear zone appears to have been situated near what is now the northern extremity of the complex.

Several of the older episodes of wrench-type shearing were related to the early evolution of what has been recently

Table 1. Sequence of tectonic events in the South Lake Igneous Complex. Dyking events 2, 3 and 4 may represent initial, main and waning phases, respectively, of the same event. Dyking events 2 and/or 4 may or may not relate to dyking events in the Wild Bight Group along the eastern boundary with the South Lake Igneous Complex

MEMBER OF COMPLEX		TECTONIC EVENTS
	DYKING EVENT 4	*(13) sporadic intrusion of mafic dykes and sills*
		(12) localized deformation of quartz diorite and possibly blue-quartz tonalite and Wild Bight Group
		(11) alteration of quartz diorite and possibly hornblende diorite
		(10) incorporation of mylonite that had a mafic protolith
QUARTZ DIORITE		*(9) intrusion of quartz-diorite bodies along and across the intrusive margins of hornblende-diorite bodies*
QUARTZ TONALITE	DYKING EVENT 3	*(8) intrusion of blue quartz-tonalite bodies and inclusion of paragneiss xenoliths. Accompanied emplacement and shearing of orthogonal suites of northeast and northwest mafic dykes (DYKE SWARM 2). Contemporaneous deformation of quartz tonalite and possibly hornblende diorite*
	DYKING EVENT 2	*(7) localized intrusion of composite mafic – felsic dykes*
		(6) restricted development of northeast and northwest shear zones in regionally undeformed bodies of hornblende diorite
		(5) incorporation of metagabbro enclave and exotic xenoliths of altered felsic tectonites
HORNBLENDE DIORITE		*(4) intrusion of hornblende-diorite bodies into northeast shear zones that rework dyke swarm 1*
	DYKING EVENT 1	*(3) emplacement and shearing of sheeted mafic dykes (DYKE SWARM 1), contemporaneous with development of northwest shear zones in metagabbro*
META-GABBRO		*(2) mylonitization and production of northeast structural grain in flaser gabbro and leucogabbro*
		(1) formation of primary cumulates in ophiolitic gabbro

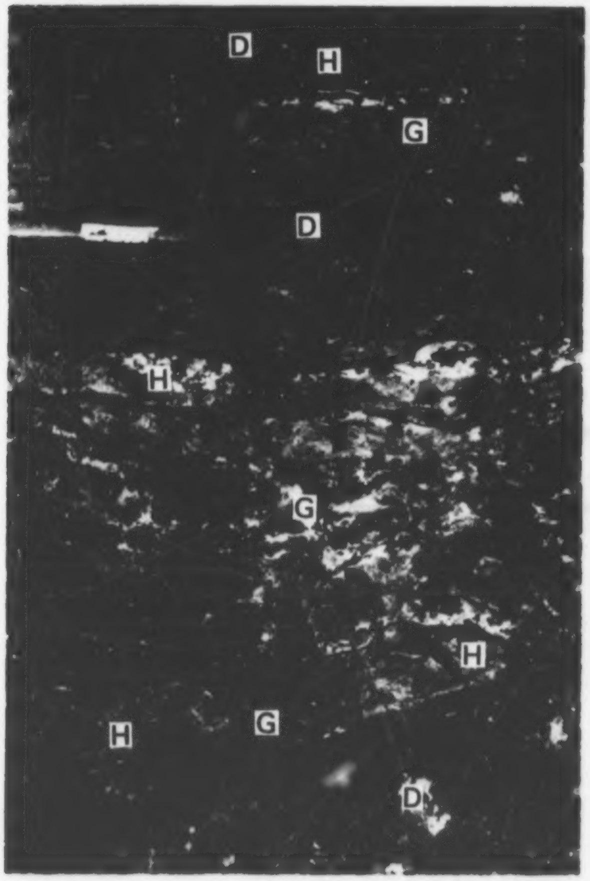


Plate 2. *Injection and crystallization of late magmatic fluids, represented by hornblendite veinlets (H), preceeded mafic dyke (D) intrusion but was, nevertheless, synkinematic with respect to emplacement of sheeted mafic dykes in metagabbro (G). The hornblendite veinlets are interpreted to occupy pinnate tensile fractures, whereas the variably foliated mafic dykes are thought to mark the location of linked shear fractures. Dextral wrench-related structures possibly controlled these discordant members of the oldest dyke swarm. Looking down on a subhorizontal erosion surface at subvertically dipping, west- and northwest-trending contacts of composite dykes,* en echelon *veinlets and altered metagabbroic host rock.*

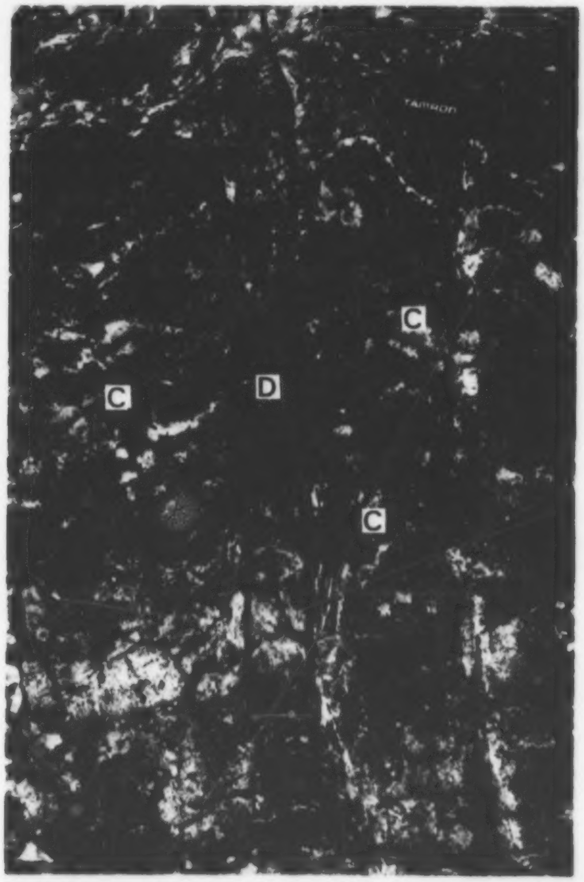


Plate 3. *An internal, wall-to-wall foliation (open symbol) in a discordant, sheared mafic dyke (D) is parallel to a host-gabbro foliation (closed symbol), which is axial planar to minor folds of primary (cumulate) layering (C). Note that the apical angle of such folds decreases toward the dyke margin, the site of a small ductile shear zone. Restricted to zones of dyke intrusion and contemperaneous with emplacement of the oldest swarm, the foliation illustrated is perpendicular to the mylonitic foliation in adjacent tracts of flaser gabbro.*

deemed the complex's ophiolitic tract (Table 1). Such deformation produced zones of flaser gabbro and strongly foliated gabbro separated by zones displaying primary igneous texture, and may have occurred entirely within Ordovician oceanic crust. It imparted, to the ophiolitic tract, a variably developed, north- to northeast-trending structural grain (see also Lorenz and Fountain, 1982). A somewhat later episode of shearing in metagabbro was associated with the emplacement of the oldest mafic dyke swarm in the complex and resulted in the development of an internal system of west- to northwest-trending shear zones. Less abundant, northeast-trending shear zones, also apparently confined to the ophiolitic tract, reworked earlier structures and the originally discordant dykes of the oldest northwest-trending swarm.

A mesozonal shear zone, whose effects were mostly manifested in the magmatic arc tract in the north of the complex, nucleated in what is now a rhombohedral enclave of ophiolitic metagabbro. There, the regional partitioning of hornblende-diorite and quartz-tonalite bodies is strikingly evident. However, in the southern parts of the complex, their relative disposition and local outcrop patterns are much more irregular and the mutual relations of diorite and tonalite are less readily discernible. It seems that sheet intrusions in the southern parts of the complex have had their emplacement less controlled by pre-existing and contemporaneous structures than those in the north. Lorenz and Fountain (1982) regarded the magmatic arc tract as a petrogenetically related, subduction-generated plutonic suite and hypothesized that

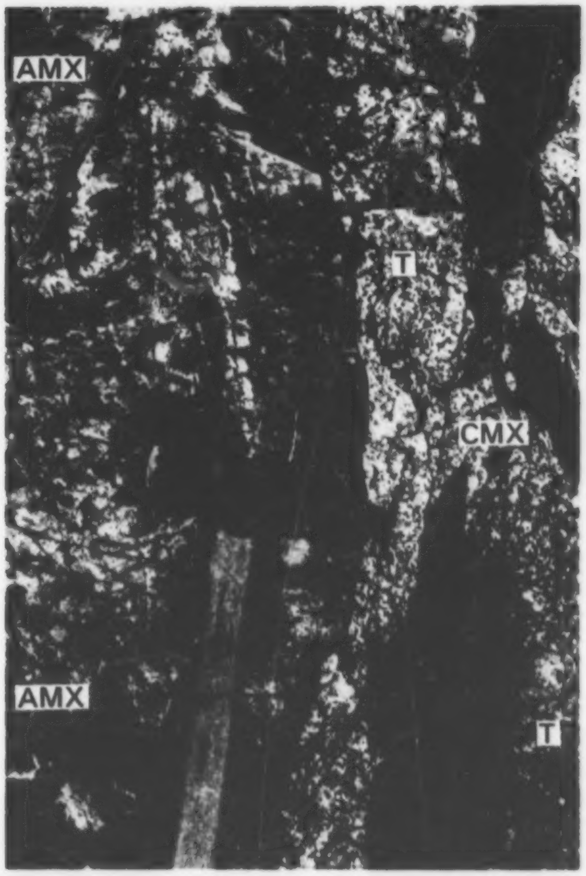


Plate 4. *Weakly foliated, blue-quartz tonalite (T) of the South Lake Igneous Complex contains cognate mafic xenoliths (CMX) and accidental paragneissic xenoliths (APX). Note the folded metamorphic banding and migmatitic segregation lits within the xenolith of paragneiss. An exotic lithology uncharacteristic of adjacent rock groups, paragneiss is also unrecognized as an internal unit of the igneous complex at present erosional level.*

these plutons originated in a marginal basin and formed below an ensimatic volcanic island arc.

The ductile shear zone hosted by the South Lake Igneous Complex had a movement history revealed, in large part, by mafic dykes that were syntectonically emplaced during four, geologically separable events (Table 1). The shear zone evolved within, and is restricted to, the complex. As hard-rock deformation proceeded, shearing was continuously focused into progressively younger intrusions, although whether synplutonic deformation inhomogeneously affected older rocks in different parts of the complex is unknown. Episodes of mesozonal synplutonic tectonism are bracketed by the ages of the ophiolitic metagabbro and later dioritic and tonalitic plutons. Obliquely oriented to and transected by the generally north-trending faults defining the complex's external boundaries, the shear zone overlapped, with declining influence, the emplacement of the intrusions comprising the South Lake Igneous Complex.

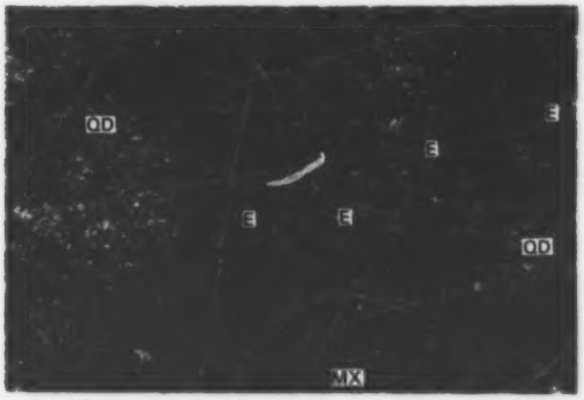


Plate 5. *Unfoliated, well-jointed, quartz diorite (QD) is host to an accidental xenolith of mylonite (MX). Along part of the external boundary of the xenolith, quartz diorite is seen to crosscut the mylonitic foliation. Note the presence of foliation-parallel, epidote-rich layers (E) within the xenolith, possibly suggesting that the protolith of the mylonite was mafic in composition.*

RELATIONSHIPS EXTERNAL TO THE COMPLEX

The external relationships that the Lower–Middle Ordovician South Lake Igneous Complex have to adjacent rocks of overlapping and younger age are exceedingly variable and notably cryptic. Regionally, the complex is located along a major structural culmination within the Middle Ordovician Wild Bight Group; however, through the particular tract of ground shown in Figure 2, it is also tectonically juxtaposed against the Upper Ordovician Point Leamington greywacke. In this region, mafic dykes and sills are ubiquitous in the South Lake Igneous Complex, common in the Wild Bight Group and rare in the Point Leamington greywacke. Generally, within each of these units, the mafic intrusions comprise texturally and compositionally distinct suites, which display variable relations with the local structures found in their host rocks.

Fossils from the Wild Bight Group and Point Leamington greywacke, recently discovered near the South Lake Igneous Complex, have been tentatively identified by S.H. Williams (personal communication, 1991). The youngest Wild Bight formation (Unit 6), which is juxtaposed against the South Lake Igneous Complex, contains graptolites indicating an *Nemagraptus gracilis* Zone age (earliest Caradoc). The oldest known Point Leamington unit so disposed belongs to the *Dicranograptus clingani* Zone (latest Caradoc). The Lawrence Harbour shale of mainly Caradoc age, although present to the immediate north and south of this region, is notably absent adjacent to the South Lake Igneous Complex.

POSSIBLE SILURIAN RELATIONSHIPS OF THE SOUTH LAKE IGNEOUS COMPLEX

The South Lake Igneous Complex is externally bounded by generally steeply east- and west-dipping faults that are openly folded by northeast-trending regional folds (Figure 2). The bounding faults apparently coalesce near Cramp Crazy Lake, where the Wild Bight Group and the Point Leamington greywacke intervene between two separate outcrop areas of the complex. From observation of minor structures developed in adjacent Ordovician stratified units, it is concluded that these faults formed as the Ordovician beds were overturned, before or during the imposition of the regional slaty cleavage. Fold asymmetry and axial-surface inclination, together with the dip direction of attendant slaty cleavage, indicate that the sharply defined, bounding structures are brittle-ductile, high-angle reverse faults or, more rarely, thrust faults (see O'Brien, 1991c for fuller treatment). Because the South Lake Igneous Complex is seen to terminate within the Wild Bight Group, it is considered to occupy a horse in these stratified rocks rather than a klippe lying above such strata (Figure 2).

Although local age control on the complex's bounding faults is currently unavailable, any regional deformation of the South Lake Igneous Complex that also affects the Point Leamington greywacke must be younger than mid-Ashgill (Williams *et al.*, *this volume*). In the Bay of Exploits area, where structures are thought to be correlative and contemporaneous with those discussed above, thrust-related folds and cleavage affect mid-Llandovery rocks as well as the Ashgill Point Leamington greywacke, and are truncated by Long Island quartz monzonite yielding a K–Ar biotite age of 448 ± 10 Ma (Mandville, 1991).

The southernmost portion of the fault along the eastern margin of the South Lake Igneous Complex juxtaposes Point Leamington and Wild Bight strata that define a regional structural domain governed by northwest-trending slaty cleavage, major periclinal folds and high-angle reverse faults (Figure 2). Although this pattern could be interpreted to result solely from coaxially refolding structures that originally trended northeast, most units in the Exploits Subzone occupy such northwest-trending domains, which are known to occur regionally throughout north-central Newfoundland.

Some reported explanations of this phenomenon are that the northwest-trending domains are medial Ordovician to Silurian in age and are overprinted by northeast-trending domains of Late Silurian age (Blewett, 1991) or, alternatively, that the older northwest-trending domains were initiated during sedimentation (Helwig, 1967) of the Middle Ordovician and older (?) parts of the Wild Bight and Exploits groups. Another possibility is that the northeast- and northwest-trending structures mutually overprinted each other, with domains developing contemporaneously on a regional scale (O'Brien, 1991c). In this explanation, the grain of these Late Silurian (?) structures, particularly the northwest set, would be inherited from and initially controlled by the structural architecture (Thon, 1980; Glenn, 1985) of underlying rocks (e.g., the steeply dipping Ordovician shear zones in the South Lake Igneous Complex).

The faulted contact of the South Lake Igneous Complex crosscuts internal stratigraphic boundaries within the Middle to Upper Ordovician part of the Wild Bight Group, the latter also being locally thrust imbricated between the South Lake Igneous Complex and the Point Leamington greywacke (Figure 2). However, if previous assumptions about the age of the South Lake Igneous Complex are correct, then the complex is too old to have originally intruded the Point Leamington greywacke. This means that either post-Late Ordovician fault displacement is large and the whole igneous complex is allochthonous, or that the South Lake Igneous Complex and the Wild Bight Group are essentially parautochthonous relative to the Point Leamington greywacke and any younger intrusive rocks. The latter scenario may be supported by a preserved interbedded relationship between Unit 6 of the Wild Bight Group and Unit 7 of the Point Leamington greywacke (cf. Helwig, 1967), a feature only observed near the South Lake Igneous Complex (see section on Relationships of Unit 3).

As a generalization, dynamothermal metamorphism that accompanied deformation in the complex's internal system of shear zones occurred at much higher grades than that which was contemporaneous with the external system of bounding brittle-ductile faults. In addition, mylonite zones are far more common within the complex than outside of it. Therefore, while it is possible that differential movement of certain units of the South Lake Igneous Complex was contemporaneous with the development of its bounding faults, any displacements producing significant unit offset could have equally well occurred during the complex's amalgamation or, later, as the complex was exhumed. However, if displacements are not large and none of the complex's units are allochthonous relative to each other or adjacent stratified units, then the original boundary between the South Lake Igneous Complex and the Point Leamington greywacke would have been necessarily stratigraphical.

POSSIBLE ORDOVICIAN RELATIONSHIPS OF THE SOUTH LAKE IGNEOUS COMPLEX

Relationships with Wild Bight Intrusions

Though regionally developed throughout the group, minor intrusions of mafic composition are locally concentrated in the Wild Bight Group to the immediate west of the South Lake Igneous Complex. There, in Unit 5 of the Wild Bight Group (Figure 2), sills generally trend northwest and dykes northeast. Some of these rocks, traditionally held to be of medial Ordovician age, intrude coarse-grained, felsic pyroclastic deposits and show irregular diffuse margins indicative of a penecontemporaneous synvolcanic origin. Other mafic dykes and sills were intruded after sericitic and silicic alteration of these Wild Bight tuffs and apparently postdated foliation development in the group, at least in this location. It is possible that some of these minor intrusions are also present in quartz diorite (Unit 4) of the South Lake Igneous Complex (Table 1).

Relations with Wild Bight Strata

Dean (1977) stated that certain members of the presumed medial Ordovician and older South Lake Igneous Complex intruded the Wild Bight Group. This relationship is not observed along the margins of any of the regionally mappable bodies depicted in Figure 2, although xenoliths with pre-incorporation fabrics in Unit 4 could be argued to belong to the adjacent Wild Bight Group. However, taken together with the relations of the synvolcanic or syntectonic dykes in Unit 5 and the conformity of the Unit 5–Unit 6 succession, this would imply Ordovician deformation of a Wild Bight volcanosedimentary sequence at a time when, according to abundant fossil evidence, most of the Exploits Subzone was being blanketed by anoxic pelagic sediment (i.e., Lawrence Harbour shale and equivalents). Until it can be demonstrated that deformation of this age affects stratified rocks in the region, it should not be assumed that any South Lake Igneous Complex unit has intrusive relationships with the Wild Bight Group.

Relations with Point Leamington Strata

One of the most critical external relationships of the South Lake Igneous Complex is the one it possibly once had with Ashgillian strata of the Point Leamington greywacke. Laterally discontinuous lenticles of polymictic conglomerate and pebbly wacke (Figure 2) are characteristic of a variably thick succession that records the interval between the *Dicranograptus clingani* and *Dicellograptus complanatus* graptolite biozones around most, if not all, of the Point Leamington basin margin (Williams *et al.*, *this volume*). In most places, the conglomerate-wacke sequence is underlain and overlain by thin-bedded sandstone–argillite sequences in the lowest part of the Point Leamington succession. Conglomeratic and pebbly flysch lenticles cropping out near the eastern margin of the South Lake Igneous Complex have complexly interlinked protrusions or tongue-like splits, presumably related to events that caused the stratigraphic wedging of the host turbidite succession. In those parts of the Point Leamington basin that are not adjacent to the South Lake Igneous Complex, black anoxic mudstone of the Lawrence Harbour shale forms the substrate to a similar Ashgillian succession.

Relationships of Units 8 and 9

Pebbly wacke-polymictic conglomerate sequences on Western Arm Brook and near South Lake comprise two separate bodies of Unit 9, both of which directly overlie Unit 8 grey shale of the Point Leamington greywacke (Figure 2). In both localities, the thin grey shale horizons underlying Unit 9 wacke contain probable *Dicranograptus clingani* Zone graptolite assemblages. The underlying grey shales are, however, lithostratigraphically distinct entities. The grey shale lenticle on Western Arm Brook is intercalated with Unit 7 of the Point Leamington greywacke, whereas the lenticle near South Lake gradationally overlies *Nemagraptus gracilis* Zone strata belonging to Unit 6 of the Wild Bight Group. Unit 8 shale and Unit 9 wacke of the Point Leamington greywacke are locally separated by a shale-matrix olistostrome that contains large blocks of coral-bearing limestone and celadonite-bearing chert at its base. Some of the transported material in the debris flow located on Western Arm Brook is also found *in situ*, and as reworked clasts, in Unit 6 of the Wild Bight Group. Only short distances along strike from the olistostrome, where neither *D. clingani* shale nor pre-*D. complanatus* conglomerate form part of the Upper Ordovician flysch, splits or wedges within the host turbidite sandstone succession are not observable.

Relationships of Unit 3

In several locations, polymictic conglomerate lenticles of the Point Leamington greywacke are in direct contact with Unit 2 diorite and Unit 3 tonalite of the South Lake Igneous Complex (Figure 2). On Martin Lake, Point Leamington wackes (Unit 9), lying tectonically adjacent to Unit 3, contain abundant, euhedral, blue-quartz clasts and rare prisms of fresh hornblende in addition to pebbles of tonalite, other granitoids, felsic volcanic rocks and celadonitic chert. This implies insignificant chemical weathering of the clasts' source area and little mechanical breakdown of the wackes' detrital constituents. A nearby provenance is likely; the adjacent body of blue-quartz tonalite apparently being preferentially eroded and locally transported. Furthermore, these observations suggest that the South Lake Igneous Complex and the Wild Bight Group were already in close proximity to each other by late Caradoc time, with both supplying detritus to the Point Leamington sedimentary basin. Onlap of the stratigraphically lowest conglomeratic sequence or uplift of this source area, along the basin margin immediately prior to the beginning of the *Dicellograptus complanatus* interval, may have resulted in Unit 9 wacke being originally deposited on Unit 3 of the igneous complex. Alternatively, either a locally anomalous amount of tectonic removal of underlying strata (Unit 7) has occurred where Unit 9 is in direct contact with Unit 3, or the northwest-trending pericline affecting Units 7 and 9 formed prior to the northeast-trending fault bounding Unit 3 (Figure 2).

Summary of Late Ordovician Interrelationships

Deposition of the stratigraphically lowest, 'feather-edged' lenticles of conglomerate and wacke was apparently triggered by localized syn-sedimentary uplift during an interval overlapping the Caradoc–Ashgill boundary. At least one of the accumulation sites of polymictic conglomerate was initially controlled by gravity-driven slumping and mass wasting within a specific submarine channel, peculiar to one of the margins of the Point Leamington basin.

Ashgillian uplift and resedimentation of Caradoc black mudstone occurred during the *Dicellograptus complanatus* and *Dicellograptus anceps* intervals (e.g., Williams *et al.*, *this volume*); it affected those parts of the basin underlain by the Lawrence Harbour shale. As discussed above, uplift of that portion of the basin margin, which abutts the South Lake Igneous Complex, occurred earlier. Apparently, such uplift and related erosion also affected hemipelagic chert-covered ridges (Unit 6), only partially blanketed by a more oxidized, grey mudstone facies. Locally, seamounts composed of South Lake tonalite (Unit 3) and adjacent Wild Bight volcanic rocks (Unit 5) may have been sampled. Although such tectonic uplift was aerially restricted, the phenomenon may be regionally significant. Olistrostrome-related erosion and scouring of gullies in Caradoc mudstone is also seen at the base of an Ashgill–Llandovery hemipelagic succession on Upper Black Island, where the Late Ordovician–Early Silurian flysch is regionally juxtaposed against the Point Leamington greywacke (O'Brien, 1991c).

DISCUSSION

Distinct mesozonal phases of Ordovician mafic dyking occurred during the tectonomagmatic evolution of the South Lake Igneous Complex (Table 1). Within the dyke rocks and plutons of this complex (Lorenz and Fountain, 1982), and within the dyke rocks and volcanic rocks of the Wild Bight Group (Swinden *et al.*, 1990), important Ordovician episodes of post island-arc, rift-related, tholeiitic-alkalic magmatism have been recognized through petrogenetic analyses. In both lithostratigraphic units, such magmatism was interpreted in terms of a spreading back-arc basin model, although precise age constraints on the rift types of magmatism were unavailable to any of the above workers.

Any temporal or spatial linkage between late Wild Bight volcanism and late South Lake plutonism must await an integrated geochronological, paleontological and allied petrochemical study. However, based on the data presented in this paper, one might provisionally hypothesize that both units were exhumed, uplifted and in proximity to each other by the early Ashgill, prior to much of the infilling of the Point Leamington sedimentary basin. As previously suggested, related tectonism may have affected several formations of the Exploits Group underlying the lower Middle Ordovician Strong Island chert (O'Brien, 1991b,c), with faulting associated with this regionally extensive phase of tectonic movement being locally focused in the New Bay Fault Zone (Williams *et al.*, *this volume*). In this regard, it is noteworthy that dyke swarm-related fracture patterns in ophiolite-spreading centres have been reported to control the future site of olistostrome-filled, paleotopographic valleys in oceanic basement at evolving ocean-continent transform margins (Harper *et al.*, 1985; Arnott *et al.*, 1985).

In the area examined, submarine canyon-fill deposits of Late Ordovician age are interpreted to have been carved into small, fault-bounded basins of upward-coarsening flysch, which lay adjacent to exhumed Iapetan tracts of Early Ordovician ophiolitic and magmatic arc rocks. Such a scenario is also envisaged for similar aged rocks in the Appalachian–Caledonian foldbelt of Norway (Minsaas and Sturt, 1985). There, subaerially exposed, arc-ophiolite terrains formed uplifted ranges on the flanks of a Late Ordovician–Early Silurian marginal basin, which incorporated numerous, partially isolated depocenters. These Norwegian mid-Paleozoic sediments are thought to provide evidence of renewed, post-obduction rifting of Iapetan ophiolites and magmatic arcs, previously accreted to the continental margin of Baltica (Thon, 1985). In north-central Newfoundland, the Middle to Upper Ordovician rocks in Unit 6 of the Wild Bight Group (partial equivalent of the Strong Island chert ?) may represent the oldest part of an analogous, if not correlative, overstep sequence seaward of the Gondwanan continental margin.

CONCLUSIONS

As first suggested by Lorenz and Fountain (1982), the South Lake Igneous Complex records the mid-crustal evolution of a Lower–Middle Ordovician magmatic arc emplaced into a probable ophiolitic relic of the Dunnage Zone. During the rise of the arc magma in precursor and contemporaneous shear zones, sialic material, which appears to be exotic relative to the preserved enclave of oceanic crust, was entrapped in each successive member of the igneous complex. Its relationships with Ordovician stratified rocks of overlapping and younger ages in the Exploits Subzone are complicated by later tectonism of presumed Late Silurian age. Unfaulted contacts between plutonic rocks of the South Lake Igneous Complex and volcanic rocks of the Wild Bight Group were not observed, although both may have shared common mafic dyking episodes during the initial phase of uplift and excumation of the magmatic arc, and during subsequent regional deformation. A siliciclastic flysch of uppermost Caradoc and Ashgill age (Point Leamington greywacke) is postulated to have originally infilled a small, fault-controlled, cover basin (cf., Arnott *et al.*, 1985), whose local basement included the South Lake Igneous Complex and the Wild Bight Group.

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GEOLOGICAL DEVELOPMENT OF THE AVALON ZONE, THE EASTERNMOST GANDER ZONE, AND THE DUCTILE DOVER FAULT IN THE GLOVERTOWN (NTS 2D/9, EAST HALF) MAP AREA, EASTERN NEWFOUNDLAND

S.J. O'Brien and R.E. Holdsworth[1]
Newfoundland Mapping Section

ABSTRACT

The eastern portion of the Glovertown (NTS 2D/9) map area comprises two contrasting domains, each characterized by profoundly different geological histories. Rocks in the east form part of the Avalon Zone, and include strata previously assigned to the late Precambrian Love Cove and Musgravetown groups. The Avalonian geological record in this area is of widespread shallow-marine to terrestrial sedimentation and volcanism, and subsequent low-grade regional metamorphism accompanying a relatively simple, slate-belt-style of deformation. In contrast, the Gander Zone to the west is represented by a sillimanite-grade gneiss complex (Hare Bay Gneiss) that is host to numerous granitic sheets, many of which are of regionally mappable dimensions. Almost all have been deformed to varying degrees with the surrounding gneisses.

Adjacent to their north-northeast-trending boundary, the Avalon and Gander zone rocks are affected by a zone of early ductile sinistral transpression, which is overprinted by lower grade ductile and subsequent brittle, dextral displacements. The early ductile deformational event in the Gander Zone is associated with a kilometre-scale, steep belt of focussed high-grade metamorphism, migmatization and granitic plutonism. In the northern half of the map area, an almost complete section of the later ductile, dextral mylonites is preserved within the ductile Dover Fault Shear Zone. In the southern part of the area, however, these mylonites are cut out by later dextral faults. The coarse-grained, K-feldspar porphyritic Terra Nova Granite postdates all but the youngest brittle movements along the Gander–Avalon zone boundary and in the adjacent country rocks.

INTRODUCTION

The eastern part of the Glovertown (NTS 2D/9) map area (Figures 1 and 2) is traversed by a complex, north-northeast-trending, steeply dipping tectonic zone, characterized by a reactivated transpressional deformational history. In the study area, as is the case elsewhere in the northeastern Newfoundland Appalachians (e.g., Blackwood, 1978; Holdsworth, 1991), this tectonic zone juxtaposes a lower Paleozoic, amphibolite-facies, granite–gneiss terrane to the west (Gander Zone; Williams, 1979) with a late Precambrian, greenschist-facies, terrestrial to shallow-marine, volcano-sedimentary terrane to the east (Avalon Zone; Williams, 1979). Evidence of any linkage between rocks on either side of this boundary prior to their initial juxtaposition is not apparent. Both regions, however, do share a common structural history related to movements along their mutual boundary. A broad zone of early ductile, sinistral transpression in both the Avalon and Gander zones is overprinted by lower grade ductile dextral displacements that form the boundary observed at the present day–the Dover Fault (Younce, 1970; Blackwood and Kennedy, 1975). Subsequent brittle, dextral faults disrupt and crosscut earlier ductile structures, including the Dover Fault. In the eastern

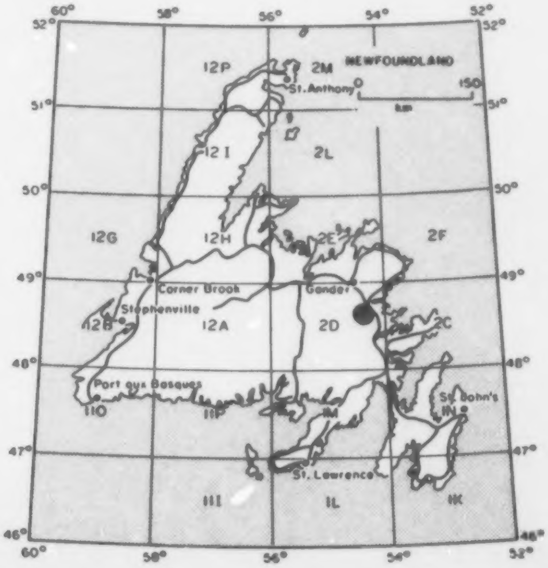

Figure 1. *Approximate location of the study area.*

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[1] Department of Geological Sciences, University of Durham, Durham, U.K.

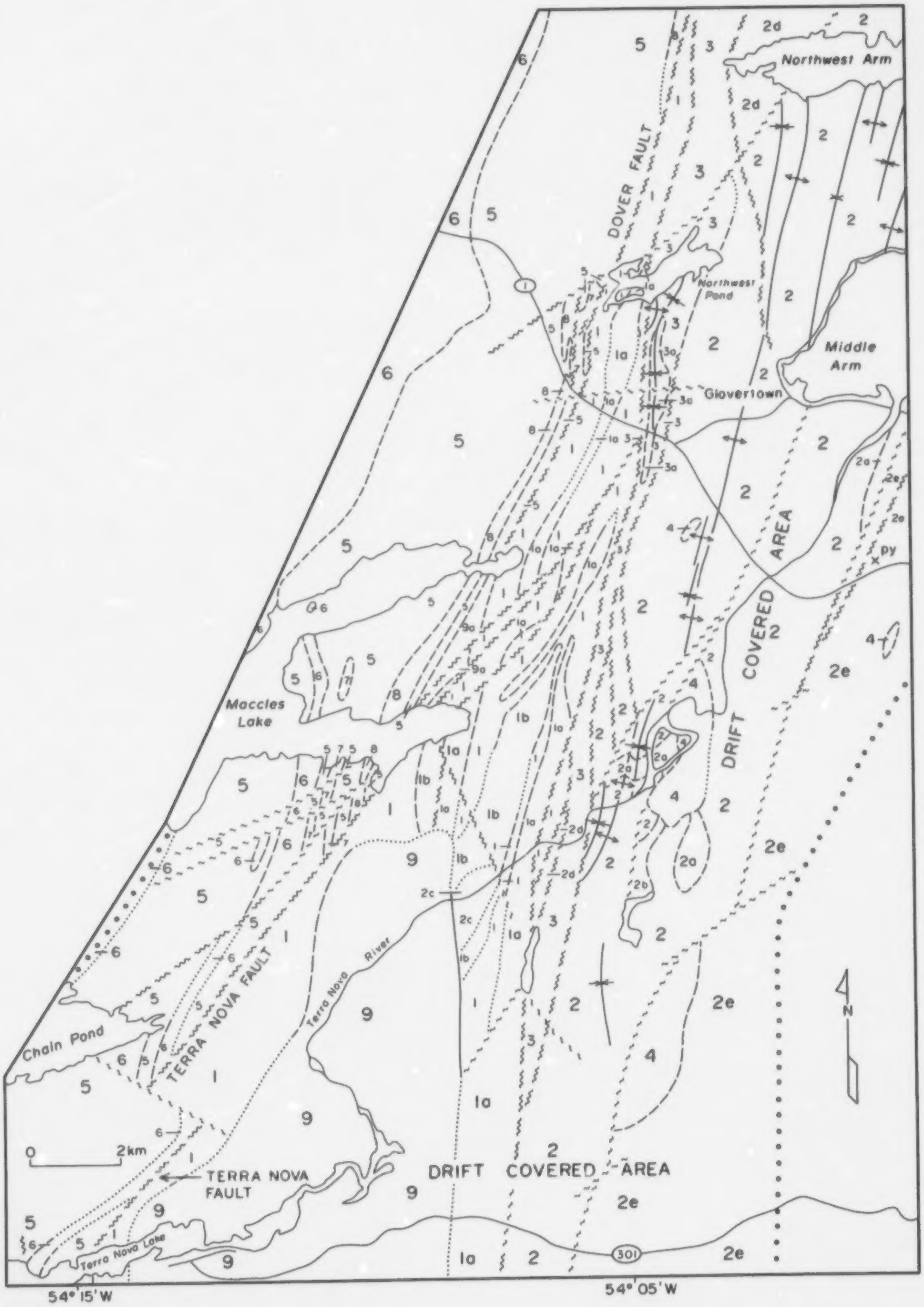

Northwest Arm
DOVER FAULT
Northwest Pond
Middle Arm
Glovertown
DRIFT COVERED AREA
Maccles Lake
Chain Pond
TERRA NOVA FAULT
Terra Nova River
Terra Nova Lake
DRIFT COVERED AREA
0
2km
N
py
1
301
54° 15'W
54° 05'W

Figure 2. *Preliminary geological map of the eastern part of the Glovertown map area.*

LEGEND (for Figure 2)

DEVONIAN?

TERRA NOVA GRANITE

9 *Massive to very weakly foliated, pink, medium- to mainly coarse-grained, K-feldspar porphyritic, biotite ± hornblende granite; rare aplite;* **9a**: *coarse-grained, quartz-rich, biotite granite; minor pegmatite and graphic granite*

SILURIAN?

DOVER FAULT GRANITE (UNIT 8)

8 *Mylonitic, variably megacrystic to coarse-grained, K-feldspar porphyritic granite; ultramylonite; amphibolite; minor, weakly foliated granite*

7 *Foliated, fine-grained muscovite–biotite leucogranite; locally contains unseparated megacrystic granite sheets*

6 *Foliated, pink, grey and white, K-feldspar megacrystic biotite granite, locally intruded by minor unseparated leucogranite; contains minor metasedimentary rocks*

EARLY ORDOVICIAN OR EARLIER

HARE BAY GNEISS

5 *Unseparated metasedimentary gneiss and amphibolite, mobilized gneiss and orthogneiss*

LATE PRECAMBRIAN

4 *Unseparated gabbro and diabase*

MUSGRAVETOWN GROUP

3 CROWN HILL FORMATION: *red, maroon and purple, planar-bedded and crossbedded sandstone; minor shale and pebble conglomerate;* **3a**: *volcanic breccia, aquagene tuff, red and green sandstone*

2 ROCKY HARBOUR FORMATION: *green and yellow, crossbedded sandstone; minor green conglomerate and red sandstone;* **2a**: *massive and flow-banded rhyolite; minor felsic pyroclastic rocks;* **2b**: *basalt, mafic tuff and associated metabasic rocks;* **2c**: *highly cleaved, green sandstone and hornfels;* **2d**: *undivided, red and green sandstone;* **2e**: *variably strained, mainly green, massive, thin-bedded and crossbedded sandstone and phyllite; contains unseparated flows and dykes of mafic composition and minor felsic tuff*

1 *Moderately to highly strained green, yellow, buff and white rhyolitic to dacitic tuff and volcanic breccia; phyllonite; sericite schist;* **1a**: *mainly moderately strained, red, maroon and pink felsic volcanic rocks; parataxitically banded rhyolite, ash-flow tuff and rheoignimbrite; minor unseparated tuffaceous sedimentary rocks;* **1b**: *basalt, mafic tuff and minor chlorite schist*

SYMBOLS

Geological contact (defined, approximate, assumed)..

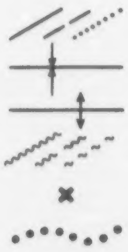
Axial trace of syncline..

Axial trace of anticline..

Fault (defined, approximate, assumed)..

Mineral occurrence..

Limit of geological mapping..



part of the Gander Zone, widespread granitic magmatism and high-grade metamorphism are associated with the early, ductile, sinistral movements.

AVALON ZONE

The Avalon Zone rocks are disposed in continuous, north- to north-northeast-trending belts that are bounded, in most cases, by faults. Adjacent to the Gander Zone, there is an extensive tract of phyllonitic to moderately or weakly cleaved, low-grade, volcanic rocks (Unit 1; Figure 2), mapped by Jenness (1963) and subsequent workers as part of the late Precambrian Love Cove Group. This exclusively volcanic belt is bounded to the east by redbeds, which in turn are underlain by clastic sedimentary rocks of shallow-marine to terrestrial origin (Rocky Harbour and Crown Hill formations; Figure 2). Both of the latter units have been assigned by Jenness (1963) to the late Precambrian Musgravetown Group. The contact of the 'Love Cove' belt with the red clastic sedimentary unit at the top of the Musgravetown Group succession is faulted. However, several lines of evidence suggest that rocks previously assigned to the Love Cove Group, in this map area, are significantly younger than the type Love Cove Group, which underlies a thick and extensive succession of deep-marine turbidites (Connecting Point Group) farther to the east (O'Brien, 1987; O'Brien and Knight, 1988). It is likely that, within the study area, much of the volcanism is, in the broadest sense, coeval with Musgravetown Group sedimentation, although their exact stratigraphic relationship remains uncertain. Hence, the name 'Love Cove Group', previously used by Jenness (1963) and subsequent workers to designate these rocks, will not be employed here.

UNIT 1: UNNAMED VOLCANIC ROCKS

The westernmost of the Avalonian stratigraphic units is composed of inhomogeneously deformed, variegated volcanic rocks that lie in a northward-thinning belt bordered by the Gander–Avalon tectonic boundary and by redbeds of the Musgravetown Group (Crown Hill Formation). These rocks form part of a much more extensive volcanic belt that extends from the coast of Bonavista Bay southward, for a distance of approximately 100 km, to the Ackley Granite (Colman-Sadd *et al.*, 1990).

The volcanic flows and pyroclastic rocks represent a continuum of compositions ranging from rhyolite to dacite and, more rarely, andesite and basalt. Texturally, they vary from phyllonitic mylonites to moderately cleaved rocks, and in some areas, may carry only a weak fracture cleavage. There appears to be a general westward, albeit non-systematic, increase in deformation toward the Gander Zone, although the regional strain is inhomogeneous. A three-fold lithological division of this belt has been established; no internal stratigraphic order is implied. It remains possible that rocks with significantly different ages occur within this westernmost belt of Avalonian rocks.

Unit 1 is mainly composed of moderately to highly strained, green, yellow, buff and white, rhyolitic to dacitic tuff. These fine-grained rocks are, in many areas, spatially associated with subordinate amounts of monolithic and polylithic volcanic breccia (Plate 1). Adjacent to the Gander–Avalon zone boundary, the tuffs and flows are reduced to green phyllonites. Narrow zones of sericitic schist derived from the above rocks are locally developed throughout this belt. Primary banding is preserved, but in many cases, it has been transposed into parallelism with a single cleavage. In some localities, the tuffs are crosscut by highly cleaved and folded mafic dykes; in general, however, dykes are not particularly widely developed within this belt. Thermal metamorphism of Unit 1 tuffs in the aureole of the Terra Nova Granite has produced extensive areas of hornfels, which in the vicinity of Maccles Lake, have largely obliterated tectonic fabrics adjacent to the boundary with the Gander Zone.

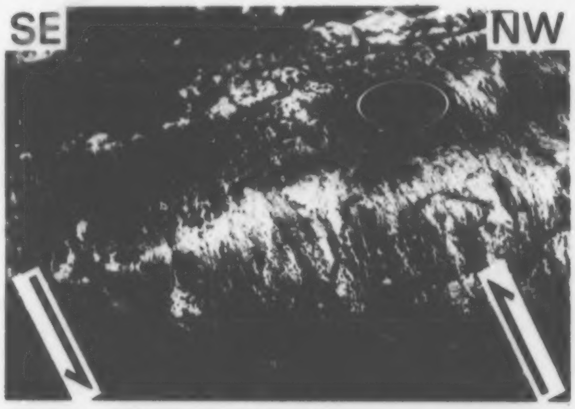


Plate 1. *Volcanic breccia of Unit 1, viewed toward the southwest, south shore of Maccles Lake (southeast arm). Note that asymmetric wrapping of the clasts in a steep surface oriented parallel to mineral lineation gives a northwest-side up sense of shear.*

A distinctive subunit (1a), consisting mainly of moderately strained, red, maroon and pink felsic volcanic rocks, comprises four semi-continuous bands that extend along much of the Unit 1 belt. A singularly significant lithological association within this subunit is parataxitically to flow-banded rhyolite, ash-flow tuff and rheoignimbrite, exposed near the southeast end of Maccles Lake (Plate 2). In that area, the latter rocks demonstrate a marked variation in strain. Over a distance of several metres, flows in which intricate, primary banding is preserved become progressively more deformed, until they are finally reduced to schists. Significantly, these texturally and compositionally distinctive volcanic rocks closely resemble the lithology most diagnostic of the Bull Arm Formation of the Musgravetown Group in the Bonavista Bay area, and in its equivalents elsewhere (O'Brien, 1987; O'Brien and Knight, 1988).

Less-common mafic volcanic rocks (subunit 1b) are mainly fine-grained and, in places, vesicular basaltic flows, which are largely confined to a northward-closing structure in the central portion of the study area. As way-up criteria are not preserved, the stratigraphic relation of these rocks

Plate 2. *Parataxitic banded rhyolite of subunit la, south shore of the southeast arm of Maccles Lake.*

to the other subunits remains equivocal. A smaller band of epidotized, mafic lithic tuff and coarser-grained, mafic breccia is faulted against subunit la rhyolite on the southeast arm of Maccles Lake.

In most previous accounts (e.g., Jenness, 1963; Blackwood, 1977; Hussey, 1979), the Unit 1 volcanic rocks and their along-strike equivalents have been included in the Love Cove Group. This unit is characteristic of the northwest Avalon Zone and traditionally is considered to occupy two separate belts separated by an intervening area of Musgravetown Group rocks. The volcanic rocks in question here, lie within the western Love Cove belt. Controversy has long surrounded the nature of this group's external contact relationships, and in particular, its contact with the Musgravetown Group. Jenness (1963) stated that the Love Cove–Musgravetown group contact was everywhere tectonic. However, on the basis of regional considerations, he assigned the Love Cove Group an older age and proposed that the two groups were separated by an angular unconformity. Younce (1970) disagreed, arguing that rocks previously assigned to the Love Cove Group throughout Bonavista Bay were, in part, deformed equivalents of the Musgravetown Group. Hussey (1979) subsequently contended that the Musgravetown Group conformably overlies the Love Cove Group, thus challenging Jenness's assertion that the Love Cove Group suffered regional deformation and metamorphism prior to Musgravetown Group deposition. Reusch and O'Driscoll (1987) inferred that the main western belt of the Love Cove Group was thrust eastward over the adjacent Musgravetown Group. Most recently, Holdsworth (1991) has presented evidence that supports a post-Musgravetown Group age for the western belt of Love Cove Group rocks in the Hare Bay–Dover area.

In the present study area, the boundaries of the unnamed Unit 1 volcanic belt are faulted. However, west of Glovertown, at Maccles Lake Brook and in the high ridges west of the Terra Nova River, the upper kilometre of the Musgravetown Group is observed to young consistently westward toward Unit 1. Moreover, significant volcanic intercalations occur in the uppermost part of the Musgravetown Group. Finally, texturally uncommon volcanic rocks, characteristic of the Musgravetown Group, are present within the Unit 1 succession. These three significant observations are consistent with the hypothesis that at least some of the Unit 1 volcanic rocks are broadly coeval with and possibly younger than some of the adjacent terrestrial sedimentary rocks of the Musgravetown Group. Until more conclusive data are available, the authors suggest that the name Love Cove Group be reserved for the volcanic rocks of demonstrable pre-Musgravetown Group age that conformably underlie the Connecting Point Group (cf. O'Brien and Knight, 1988; Knight and O'Brien, 1988). To date, such rocks have only been identified in the eastern Love Cove belt.

UNIT 2: ROCKY HARBOUR FORMATION

The most extensive Avalonian stratigraphic unit in the Glovertown area is composed of a monotonous succession of crossbedded, yellow-green sandstones, part of the Rocky Harbour Formation of the late Precambrian Musgravetown Group (Jenness, 1963). Within the study area, these sandstones pass conformably upward into the topmost red clastic sedimentary succession of that group, the Crown Hill Formation (Jenness, 1963). A few kilometres along strike to the northeast, in the adjacent St. Brendan's map area (NTS 11P/13; O'Brien and Blackwood, 1987), basal Rocky Harbour conglomerate lies disconformably upon volcanic rocks of the Bull Arm Formation.

The Rocky Harbour Formation forms structurally distinct but lithologically similar belts, separated by north-northeast-trending faults. The westernmost belt, which passes upward into the Crown Hill strata, is deformed by broad, open, regional anticlines and synclines, and is homoclinal over large areas. In contrast, rocks farther to the east appear to be characterized by generally higher, albeit inhomogeneous, bulk strain.

The principal rock types of the Rocky Harbour Formation are medium- to coarse-grained, green, yellow-green and buff, lithic sandstones. Typically, these exhibit well-defined, planar-, trough- and ripple-drift cross-stratification (Plate 3), although, in places, thin planar-bedded sandstones are predominant. The amplitude of crossbeds is variable and in many areas exceeds 50 cm. In several places, slumps and rip-ups are well preserved in these sandstones. Green pebble and cobble conglomerates, in places rich in green and, in one case, red, fine-grained felsite and flow-banded rhyolite clasts, are locally interbedded with the finer grained rocks. Subordinate amounts of red and maroon, fine-grained sandstone and, more rarely, granule conglomerate are interstratified with the other sandstones throughout the succession, most notably near the top of the formation. The red clastics are locally rich in coarse-grained, detrital muscovite.

A several-decametre-thick unit composed of interstratified felsite, rhyolite, and fine-grained felsic fragmental volcanic rocks (subunit 2a) occurs within the Rocky Harbour Formation near Grant Falls, on the Terra

Plate 3. *Crossbedded sandstones typical of the Rocky Harbour Formation, Musgravetown Group.*

Nova River. The variegated volcanic rocks overlie crossbedded sandstones, and are succeeded by thick-bedded sandstones having pebbly interlayers that are, in turn, overlain by a basalt flow and more cross-stratified sandstones. The latter rocks then persist to the top of the formation, attaining a thickness in the order of 1 km. Less continuous bands of rhyolitic flows and fine-grained, felsic tuffaceous rocks are more extensively developed in the eastern part of the Rocky Harbour Formation. Mafic plutonic rocks in the form of numerous dykes and narrow, sheet-like gabbroic bodies cut the eastern Rocky Harbour belt, whereas extrusive rocks of similar composition (subunit 2b) are rare.

At the top of the Rocky Harbour Formation, the yellow-green, crossbedded sandstone lithofacies becomes interbedded with pebbly volcanogenic sandstone and red-maroon tuffaceous sandstone. The contact with the overlying Crown Hill Formation is gradational over a few metres, and is drawn at the point in the succession where red sandstones become the predominant rock type.

UNIT 3: CROWN HILL FORMATION

A narrow but continuous band of red terrestrial clastic rocks overlies the Rocky Harbour Formation in the central portion of the study area and is mapped as part of the Crown Hill Formation of the Musgravetown Group (Jenness, 1963). The total thickness of the exposed red clastic succession in the study area is probably less than 500 m.

The lower part of this succession contains planar-bedded and poorly bedded, pale-red and bright-red sandstone, siltstone and shale. Locally, thin, buff and pale-green beds are intercalated with the redbeds. Polymictic, rounded-pebble and cobble conglomerate overlies the finer grained rocks in the northern part of the area. Their detrital assemblage includes red rhyolite, red and green dacite, red and green sandstone and shale and dark-green metabasic rocks. Some of the coarse-grained red sandstones are rich in detrital muscovite. The redbeds locally pass upward into grey-green and purple, pebbly sandstone, siltstone and mudstone that contain basaltic detritus in addition to intraformational sedimentary clasts. The latter are imbricated and, in places, display evidence of soft-sediment deformation; neptunian dykes are locally developed.

Spatially associated with the grey to purple rocks are coarse-grained, crudely bedded to massive volcanic and epiclastic breccias of probable laharic origin (subunit 3a). These fragmental rocks are composed of a green, pumice-rich, sandy matrix containing felsic volcanic blocks up to 50 cm diameter, and smaller, angular to thixotrophically deformed, intraformational rip-up clasts.

The thickness (≤ 500 m) of Crown Hill Formation preserved in this area is significantly less than that exposed farther south in the type area (e.g., Jenness, 1963) but is comparable to thicknesses measured along strike to the north in vicinity of Deer Island (O'Brien and Knight, 1988). Within the study area, the stratigraphic top of the succession is not exposed; its contact with the adjacent dominantly volcanic belt (Unit 1; Figure 2) is tectonic. However, the presence of interbedded volcanic rocks near the top of the Crown Hill Formation, coupled with the observation that the Rocky Harbour–Crown Hill succession youngs regionally westward, suggests that some portion of the adjacent volcanic belt may be coeval with and, conceivably, in part younger than these terrestrial sedimentary rocks (see above).

UNIT 4: UNNAMED MAFIC INTRUSIONS

One relatively large and several smaller mafic plutons are intrusive into the Musgravetown Group within the map area (Figure 2). These poorly exposed intrusions are lithologically variable and include fine- and medium-grained gabbro, minor diorite and diabase. These are locally fractured, foliated and epidotized and are assumed to be pretectonic intrusions.

GANDER ZONE

Sillimanite-grade paragneisses and orthogneisses (Unit 5: Figure 2), part of the Hare Bay Gneiss of Blackwood (1977), are the predominant rock types within the Gander Zone in the map area. Elongate sheets of deformed megacrystic biotite granite (Unit 6: Figure 2) intrude the Hare Bay Gneiss and are very broadly synkinematic with respect to the regional high-grade metamorphism of the gneisses; their intrusion is associated with an early phase of ductile sinistral shear along the Gander–Avalon zone boundary (Holdsworth, 1991). Smaller elongate sheets of foliated muscovite–biotite granite (Unit 7; Figure 2), locally associated with the megacrystic bodies, crosscut both gneiss and megacrystic granite. A narrow, semi-continuous sheet of medium- to coarse-grained porphyritic biotite granite (Unit 8; Figure 2), intrusive into the Hare Bay Gneiss, is situated close to and sub-parallel with the Gander–Avalon zone boundary. Unlike the older granites, this body is everywhere deformed by dextral shearing.

UNIT 5: HARE BAY GNEISS

The Hare Bay Gneiss is most accurately portrayed as a gneiss complex, composed of three main components. These are, in order of relative age, (1) metasedimentary rocks and early amphibolites, (2) variably mobilized gneisses and (3) orthogneisses (cf. Holdsworth, 1991). Boundaries between individual components are, in most cases gradational, and interbanding of these units occurs on all scales. Thus, individual sub-divisions have not been mapped on a regional scale and are not distinguished on the accompanying map (Figure 2).

The metasedimentary rocks and early amphibolites, in essence banded paragneisses (cf. Blackwood, 1978), occur as screens and rafts on all scales, intimately intermingled with zones of mobilized gneiss, orthogneiss and granite. They are mostly banded metasedimentary rocks of psammitic, semipelitic and pelitic composition. Locally, they contain very minor amounts of pale-green, impure marble, which occurs as lensoid bodies up to 1 m across. Banded amphibolites form continuous interlayers or elliptical pods from a few centimetres to several tens of metres width. Contacts with adjacent metasedimentary rocks are sharp and, although tectonic strain has obliterated any vestige of discordance, an early intrusive origin for the amphibolites seems most likely (Holdsworth, 1991).

The mobilized gneisses are, in most cases, interpreted to have formed by the pervasive partial melting of extensive regions of semipelitic metasedimentary rocks. They are typically dark grey and coarse grained, and have a well-defined granodioritic leucosome. In most areas, the mobilized gneisses contain abundant, small schlieren and enclaves, which are derived from the parent sedimentary rock. Examples of highly deformed mobilized gneisses are well exposed in outcrops along the southern shore of the northeast arm of Maccles Lake (Plate 4).

The orthogneisses are grey to grey-pink rocks of granitic to granodioritic composition. In addition, pyroxene–hornblende-bearing dioritic orthogneiss is also locally present; for example, parts of the southeast arm of Maccles Lake. Its contacts are not exposed and thus its relation to other orthogneisses are not clearly understood. Bands of dioritic orthogneiss, less than 10 m thick, are also exposed in roadcuttings along Route 1.

In most cases, the orthogneisses are not derived from mobilization of the adjacent metasediments, and they may simply represent the earliest members of the granitic intrusive suite. However, in this discussion, they are defined separately, because at least some may prove to be older bodies (see, for example, Holdsworth, 1991, p. 112). Nevertheless, an apparent gradation between orthogneiss and granite does exist on all scales.

Within the shear zone associated with the Dover Fault (see below), the Hare Bay Gneiss is reduced to discoloured, dark-green, chlorite–sericite phyllonites containing numerous feldspar porphyroclasts and variable numbers of millimetre- to centimetre-scale, pink mylonitic lenticles. The latter appear to represent intensely deformed granite veins, and their presence may be used to distinguish these phyllonites from those derived from adjacent Avalon Zone volcanic rocks.

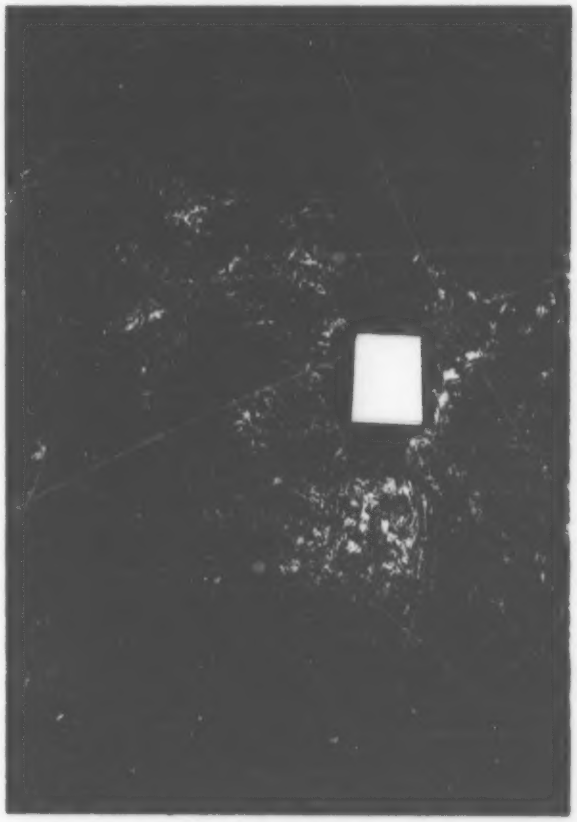


Plate 4. *Typical early ductile folds in mobilized, semi-pelitic Hare Bay Gneiss (Unit 5) outside of the Dover Fault Shear Zone, northeast arm of Maccles Lake.*

UNITS 6 and 7: MEGACRYSTIC BIOTITE AND TWO-MICA GRANITES

Megacrystic biotite granite (Unit 6; Figure 2) and intimately associated units of pink, two-mica granite (Unit 7; Figure 2) are here interpreted to represent the earliest intrusive bodies in the Gander Zone exposed within the study area. The megacrystic biotite granite intrusions are very similar to other sheets of this type observed around Bonavista Bay, most notably the Lockers Bay and Cape Freels granites (Blackwood, 1977; Jayasinghe, 1978). Like those in the study area, these latter granitic intrusions are closely associated with numerous sheets of two-mica granite.

In the study area, these granites are extensively developed within the tract of Hare Bay Gneiss south of Maccles Lake, where they form several composite, elongate, sheet intrusions. These bodies, which vary from a few centimetres to many tens of metres across, have been emplaced sub-parallel to the regional banding in the gneisses. In all cases, the presence of screens and elongate enclaves of gneiss is a common feature of the intrusions. Adjacent country rocks are often extensively and intimately sheeted with megacrystic granite on a centimetre- to metre-scale (Plate 5). A larger and more homogeneous body of foliated megacrystic biotite granite bounds the western margin of the Hare Bay Gneiss north of Maccles Lake. Near the community of Terra Nova, both megacrystic biotite and two-mica granites host large inclusions of medium- to coarse-grained diorite and gabbro. The diffuse nature of gabbro - granite contacts coupled with locally developed magma mixing textures (Plate 6) support the existence of a co-genetic mafic component to the granitic magmatism.

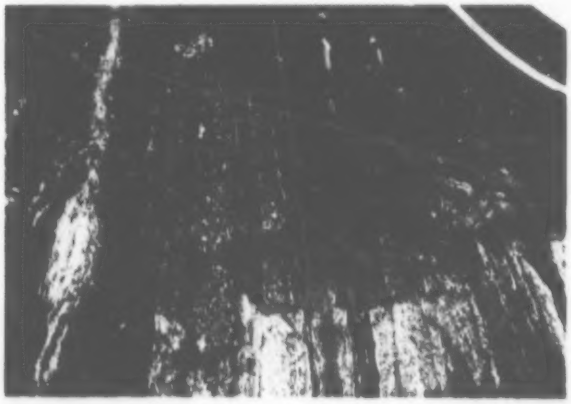


Plate 5. *Plan view of sheets of megacrystic biotite granite in metasedimentary paragneiss, Hare Bay Gneiss (Unit 5), south shore of Maccles Lake.*

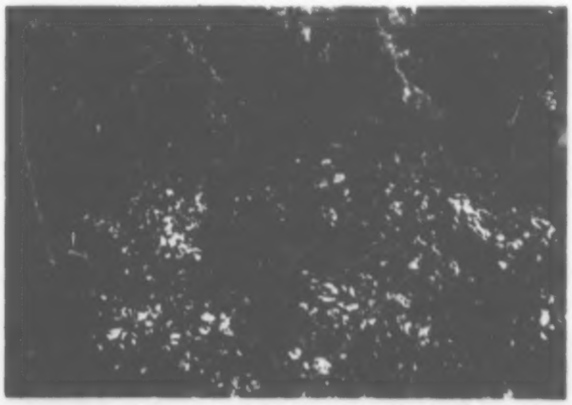


Plate 6. *Irregular and diffuse contact relationships between coexisting K-feldspar megacrystic biotite granite (light) and gabbro (dark).*

The megacrystic biotite and the two-mica granites contain strong magmatic-state foliations (cf. PFC fabrics as defined by Hutton, 1988), which are variably overprinted by solid-state fabrics. Shear-band foliations are abundant within these rocks and consistently indicate sinistral senses of displacement.

The eastern margin of what has been previously mapped as Maccles Lake Granite is also a foliated megacrystic biotite granite, associated with numerous sheets of two-mica granite. These rocks are presumably unrelated to the massive K-feldspar megacrystic granite that is typical of this pluton, suggesting that the presently defined Maccles Lake Granite is composite.

UNIT 8: DOVER FAULT GRANITE

The Dover Fault Granite is a narrow (≤500 m), elongate sheeted intrusion characterized by pink, variably megacrystic, biotite granite, which has been intensely deformed by ductile shearing associated with the adjacent ductile Dover Fault. This granite forms a semi-continuous body, well exposed between Route 1 and Maccles Lake (Figures 2 and 3); it is truncated by brittle faulting farther south. In all but one locality, the Dover Fault Granite is intruded into Hare Bay Gneiss; near Route 1, it is found in direct contact with volcanic rocks of the Avalon Zone.

Much of this intrusion is a very highly deformed mylonitic granite in which strongly wrapped feldspar porphyroclasts are the only surviving recognizable igneous component (Plate 7). A second textural type is found in the northern part of the study area where it lies immediately east of the mylonitic rocks. This second granite is characterized by weak to moderate solid-state strains and no magmatic foliation. Very rarely, narrow (<5 m) protomylonitic to mylonitic shear zones are developed. Areas of coarse-grained amphibolite and gabbro are locally associated with this low-strain granite. The faulted contact of massive and foliated rocks is exposed in a large quarry immediately north of Route 1. It is uncertain whether this less deformed granite is simply a slightly later, and therefore less-sheared body, or whether it represents low-strain regions within an otherwise strongly deformed sheeted complex.

The granite was originally emplaced as a series of sheets, sub-parallel to the regional banding in the Hare Bay Gneiss. In several areas, it is interlayered on a centimetre-scale with screens of dark, medium- to fine-grained amphibolite (Plate 8). Although these mafic units superficially resemble amphibolites of the Hare Bay Gneiss, they do not occur in the adjacent country rocks, and thus most likely represent intensely deformed coexisting mafic intrusions, emplaced synchronously with the granite.

MINOR INTRUSIONS

Minor intrusive bodies, unseparated on the accompanying map (Figure 2), include numerous sheets of pink granite, granite pegmatite and less-extensive granodiorite. These bodies, locally up to 20 m across, are found throughout the Hare Bay Gneiss, oriented parallel to, or at a low angle to, country-rock banding. At Terra Nova

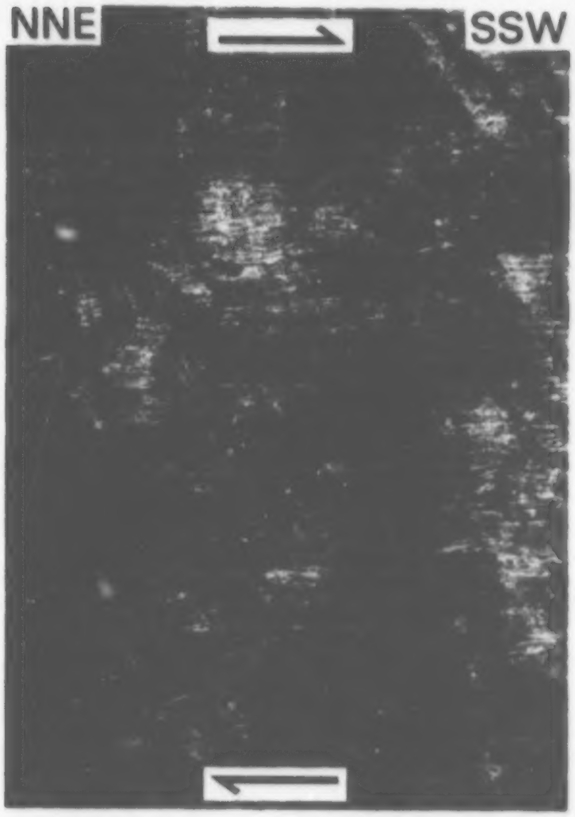


Plate 7. *The σ-porphyroclasts (feldspar) in mylonitized granite, Dover Fault Shear Zone, Maccles Lake road. Note especially well developed, dextral shear-sense indicators below and right of pencil.*

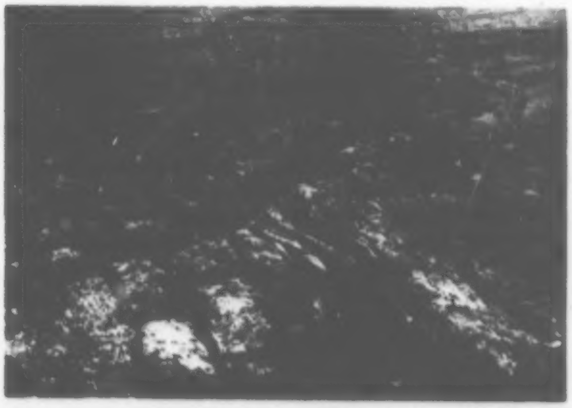


Plate 8. *Intersheeted granite, uniform dark amphibolite and metasedimentary screens, all mylonitized within the Dover Fault Shear Zone.*

Lake, the granites are cut by small dykes of unfoliated, dark grey-green diabase. Most of these are less than 1 m thick and all are crosscut by late brittle faults.

POST-DOVER FAULT INTRUSIONS

The youngest intrusive rocks in the study area, postdate all but latest brittle movements along the Gander−Avalon zone boundary and in adjacent country rocks. The largest of these is the Terra Nova Granite (Unit 9; Figure 2). This intrusion lies in the Avalon Zone portion of the study area, where it forms an irregular-shaped body with a rectilinear, north-trending eastern boundary. The diagnostic rock type of the Terra Nova Granite is pink to orange, coarse-grained, K-feldspar porphyritic, biotite−hornblende granite; a more medium-grained textural variant occurs locally. The granite is quartz-rich and in many places displays mantled feldspars, up to 5 cm in length. Although the granite is typically fresh and massive, it locally contains narrow alteration zones where a weak fracture cleavage has developed, and biotite is chloritized and feldspars are cloudy and hematitized.

The Terra Nova Granite truncates primary and tectonic layering in adjacent volcanic and sedimentary rocks. Cordierite porphyroblasts overprint the regional tectonic fabric in sedimentary rocks within its aureole. Variably cleaved hornfels locally occur as inclusions within the granite.

A probable apophysis of the main Terra Nova Granite forms a narrow and elongate, northeast-trending body (Unit 9a) situated between the northeast and southeast arms of Maccles Lake. Although petrographically similar, this granite is less texturally homogenous than the main body of Terra Nova Granite, containing coarse-grained quartz and extensive graphic or pegmatitic patches. It is mostly unfoliated, or contains a weak magmatic fabric, defined by feldspar laths aligned parallel to the regional banding. The granite is cut by small-scale brittle fractures and appears to predate very late brittle displacement along the Gander−Avalon zone boundary, along which it has been preferentially sited.

These intrusions are similar in many respects to other posttectonic granites spatially associated with the Gander−Avalon zone boundary elsewhere in Newfoundland (e.g., Ackley Granite, Dickson, 1983).

STRUCTURE OF THE GANDER−AVALON ZONE BOUNDARY

Along strike to the north, in the well-exposed region of western Bonavista Bay, three main deformational phases are recognized in the zone adjacent to the Avalon−Gander Zone boundary (Holdsworth, 1991). The earliest episode produced extensive high-strain zones, with associated sinistral sense of displacement that formed over a protracted period of time. The sinistral shear zones are viewed as broadly coeval with high-grade metamorphism, plutonism and migmatization in the eastern Gander Zone. These structures are progressively overprinted by greenschist mylonites developed within a dextral shear zone that defines the main Gander−Avalon zone boundary. Late brittle faulting disrupts this shear zone, and faults of this generation locally define the zone boundary.

In the Glovertown map area, the ductile dextral boundary is well preserved. This kilometre-wide zone of variably mylonitic and phyllonitic rocks, sited along the Gander–Avalon zone boundary, and derived from Hare Bay Gneiss and the Dover Fault Granite, is here termed the Dover Fault Shear Zone (DFSZ). From Maccles Lake northward (Figure 2), this ductile shear zone appears to be preserved more or less intact on the Gander Zone side of the boundary. South of Maccles Lake, however, a late, northeast-trending, brittle structure (the Terra Nova fault) excises a significant portion of the Gander Zone section (Figure 3), including the DFSZ. Thus, at Terra Nova Lake and Chain Pond (Figure 2), Avalonian volcanic rocks are faulted directly against gneisses and granites which, along strike and to the north, occur a considerable distance from the boundary. Within the study area, the ductile shear zone is best exposed along the northeastern and southwestern arms of Maccles Lake and also in a large quarry immediately northeast of Route 1.

DEFORMATIONAL FEATURES ASSOCIATED WITH THE DFSZ

The DFSZ is the most important and best exposed structure in the study area. For this reason, and also because its structures are directly related to movements along the Gander–Avalon zone boundary, it is described first.

Field observations indicate that the mylonites of the DFSZ formed under greenschist-facies metamorphic conditions and preserve chiefly primary, dynamic recrystallization textures. A characteristic set of minor structures have formed within the mylonites of the DFSZ, all of which indicate dextral strike-slip senses of movement (Figures 4 and 5). The same structures also overprint earlier structures in the Hare Bay Gneiss and associated granites for an additional 1 km west of the 1-km-wide zone of intense strain. The intensity of overprinting and retrogression increases progressively eastwards until most earlier structures are obliterated. It is clear that regionally, these ductile movements become increasingly focused along the Avalon–Gander zone boundary.

The DFSZ has a sub-vertical foliation, striking just east of north, upon which a ubiquitous south-plunging stretching lineation is preserved (Figures 4 and 5). Shear-band fabrics and σ-porphyroclasts consistently indicate dextral senses of shear. Open to tight curvilinear folds of the mylonitic foliation and lineation are abundant on millimetre to centimetre scale. Where axes plunge sub-vertically, folds consistently display dextral vergence (Figure 4). In addition, thin, competent horizons may develop sinistral, antithetic, brittle shears. These

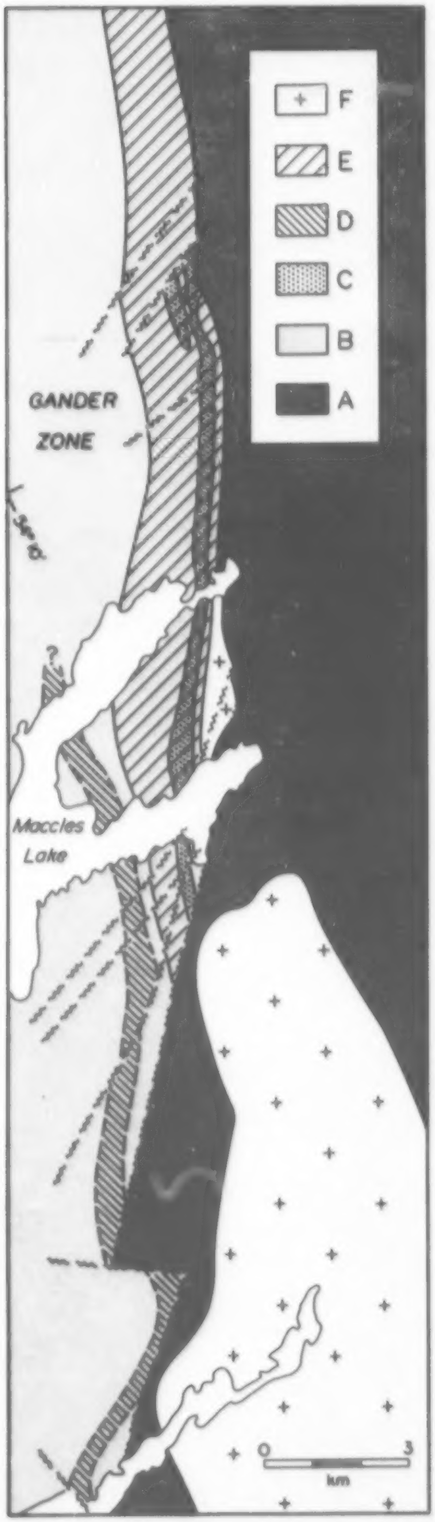
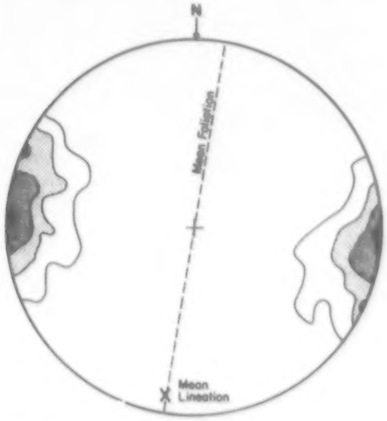
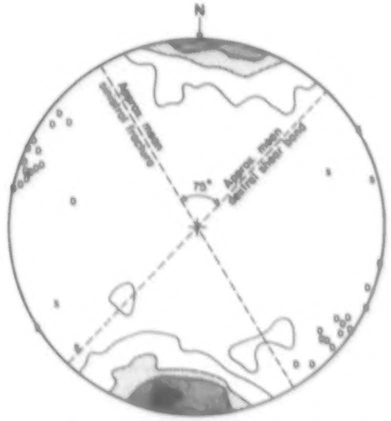


Figure 3. *Simplified map showing main structural components of the Gander–Avalon zone boundary. A = unseparated Avalon Zone; B = unseparated Gander Zone; C = Dover Fault Granite (Unit 8); D = sinistral shear zone; E = dextral shear zone; F = posttectonic Terra Nova Granite. Note that the Gander–Avalon zone boundary is the ductile Dover Fault in the northern part of the area and a later brittle fault (TNF = Terra Nova fault) in the south.*



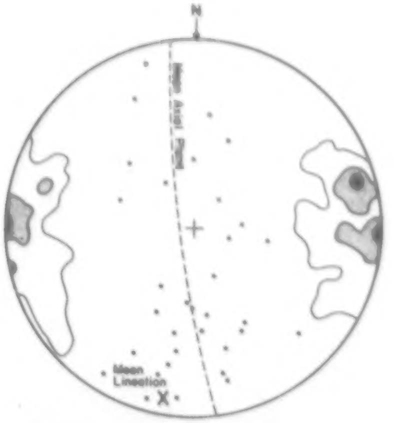
CONTOURED DATA - Poles to lithological banding (n=109) (contours - 1%, 5%, 10%, 17.5%)


N
Approx. mean sinistral fracture
Approx. mean dextral shear band
73°

CONTOURED DATA - Mineral lineations (n=84) (contours - 1%, 5%, 10%, 20%)

D SYMBOLS - Poles to dextral shear bands (n=27)

S SYMBOLS - Poles to sinistral fractures (n=5)



CONTOURED DATA - Late ductile axial planes (poles)(n=33)(contours - 1.5%, 7.5%, 16%)

Dots - Late ductile fold axes (n=34)

Figure 4. *Stereonets showing structural data from the Dover Fault Shear Zone.*

are especially common within amphibolite screens within mylonitic Unit 8 granite. In well-exposed sections of the DFSZ, numerous foliation–parallel dextral faults are exposed. These are associated with small systems of rotating fault blocks, bounded by sinistral faults, which develop in areas where two dextral faults lie close together (F in Figure 5). These faults are everywhere crosscut by abundant late brittle structures and are thought to be part of the mylonitic structural suite. They may represent local zones of brittle failure within the mylonite zone, preferentially sited along the foliation.

In exposures of Avalon Zone volcanic rocks northeast of the DFSZ, a steep north-northeast-striking, slatey cleavage containing a steeply plunging stretching lineation is preserved. The cleavage lies at low angles to banding and is verging antiform to the east-southeast. Toward the Gander–Avalon zone boundary, the strain gradually increases, so that the cleavage and banding become parallel and the lineation rotates anticlockwise toward a shallower plunge (oriented west-northwest). In regions where the lineation plunge is steep, asymmetric wrapping of volcanic clasts gives northwest-side up and sinistral senses of shear (Plate 1). With proximity to the fault, where lineations have shallow plunges, only dextral shear senses are preserved. Although it has not be demonstrated in this area, a dextral over sinistral overprint is inferred, mainly on the basis of marked similarities with the Bonavista Bay coastal exposures (cf. Holdsworth, 1991).

The geometry and dextral kinematic shear sense of the DFSZ structures, together with the greenschist-facies metamorphic grade, indicate that these structures can be directly correlated with the late ductile structures associated with the Dover Fault farther north along the west coast of Bonavista Bay (Holdsworth, 1991). In that region, the ductile shear zone is highly disrupted by later brittle faults. Thus, exposures north of Maccles Lake in the present study area are particularly important, because they preserve the late ductile Avalon–Gander Zone boundary shear zone in its most intact form.

EARLIER DUCTILE DEFORMATION

The late ductile structures of the DFSZ, described above, overprint earlier folds and gneissic fabrics within the Hare Bay Gneiss. Unfortunately, good continuous sections through these rocks are not present in the study area. At Maccles Lake, the earlier structures are tight to isoclinal folds of the gneissic banding, which plunge sub-parallel to a moderate to strong, flat-lying, north-south stretching fabric (Plate 4). These structures closely resemble the early ductile folds recognized in the Bonavista Bay region (Holdsworth, 1991), which are thought to be equivalent to the regional F_3 folds of O'Neill (1991). On the coast of Bonavista Bay, such folds are associated with sinistral shear criteria. In the study area, however, no unambiguous evidence for this shear sense was observed in those locations where overprinting late, ductile shear bands or open folds are present.

A north-trending, ca. 1-km-wide, sinistral shear zone within the Gander Zone extends from the north shore of

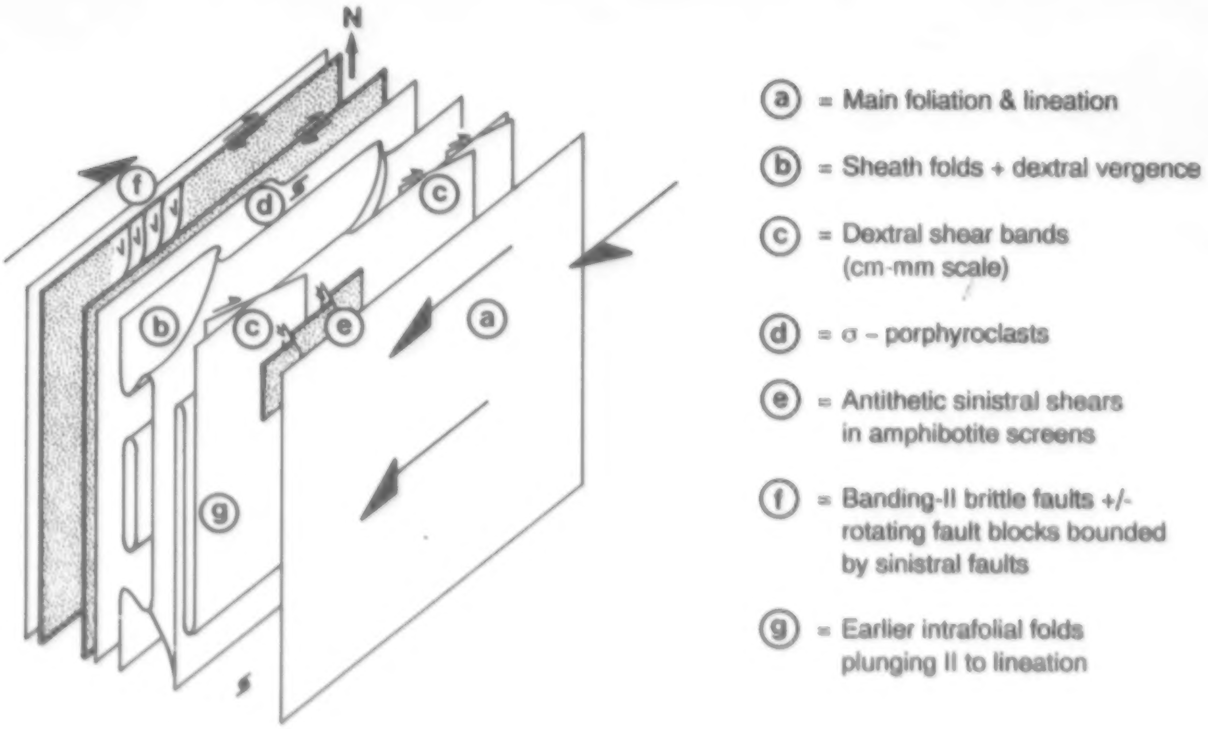

N
f
d
c
b
c
e
a
g
(a) = Main foliation & lineation
(b) = Sheath folds + dextral vergence
(c) = Dextral shear bands (cm-mm scale)
(d) = σ – porphyroclasts
(e) = Antithetic sinistral shears in amphibotite screens
(f) = Banding-II brittle faults +/- rotating fault blocks bounded by sinistral faults
(g) = Earlier intrafolial folds plunging II to lineation

Figure 5. *A 3-dimensional summary diagram of minor structures associated with the Dover Fault Shear Zone (DFSZ). Compare with data in Figure 4. Note widespread evidence for dextral shear senses.*

Maccles Lake to Terra Nova Lake (Figures 2 and 3); its boundaries are not exposed. It is characterized by locally complex, polyphase deformation, attributed to protracted, progressive sinistral shear. In this area, protomylonitic to mylonitic Hare Bay Gneiss and granites containing sinistral σ-porphyroclasts are locally overprinted by one or two phases of sinistral shear bands or millimetre- to centimetre-scale folds. These structures rework the mylonitic foliations and deform the early, north-trending, subhorizontal lineation.

The most complex part of this shear zone is exposed on the islands in Terra Nova Lake, where later sinistral folds are seen to refold earlier, sinistral, shear-band fabrics. Belts of fine-grained ultramylonite, up to 15 m across, parallel the regional foliation and in many localities are focused in fine-grained, two-mica granite sheets. Adjacent megacrystic granite units display completely different levels of strain intensity and deformational complexity. For example, a megacrystic biotite granite, containing a single weak shear-band fabric, occurs adjacent to a variably mylonitic granite of similar composition, in which two phases of sinistral folds and C−S fabrics are preserved. In such situations, the less deformed body is thought to have been emplaced at a later stage in the deformation. This view is consistent with the intrusions having been focused within the shear zone, a feature recognized elsewhere in the Gander Zone (e.g., Holdsworth, 1991). Significantly, the megacrystic granite found at the eastern contact of the Maccles Lake Granite is also affected by sinistral shear.

The relative timing of the sinistral shear-zone structures and the early ductile and late ductile events is uncertain. Whereas earlier sinistral fabrics are associated with ductile deformation of feldspars, thus implying high temperatures, later features appear to be associated with lower grade, dynamic recrystallization textures similar to those seen in the DFSZ. It is conceivable that the sinistral shear zone was active throughout both events. Significantly, its north-south trend is similar to the late ductile, Cape Freels sinistral shear zone, which also overprints a zone of earlier high-grade sinistral shear (Holdsworth, 1991).

LATER BRITTLE STRUCTURES

Small-scale, late brittle structures within the DFSZ include conjugate sets of dextral and sinistral, steeply dipping to sub-vertical faults and brittle kinks and box folds of variable plunge (Plate 9). Many faults exhibit slickenfibres, which are everywhere shallowly plunging, suggestive of mostly strike-slip displacements.

Away from larger scale, late brittle faults, minor fractures with dextral and sinistral displacements occur in roughly equal numbers and normally display mutually crosscutting relationships. This suggests that they formed as conjugate sets. As the banding, in most cases, bisects the obtuse angle between the faults, the fractures are interpreted to indicate shortening at high angles and extension sub-parallel to the regional foliation. On average, dextral faults are northeast-

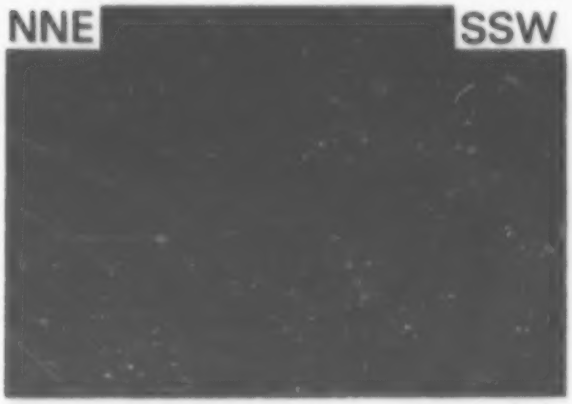


Plate 9. *Plan view of mutually crosscutting, late brittle dextral and sinistral conjugate fault sets, Terra Nova Lake.*

trending, whereas sinistral faults trend northwest. However, the considerable spread recorded in the data may be due to several orders of fault systems being present on various scales. In general, however, it appears that the conjugate faults accommodate shortening normal to the Gander–Avalon zone boundary by allowing extension along it.

The larger scale, late brittle faults are rarely exposed and, in most cases, are inferred from geological offsets or marked topographical features. Most faults trend northeast or east-northeast and have a dextral sense of displacement. The Gander–Avalon zone boundary south of Maccles Lake is marked by such late brittle faults. Clockwise changes in foliation strike (up to 50°), reflecting fault drag, may occur in zones several hundred metres either side of major dextral faults. Locally, asymmetric, Reidel-type, minor faulting patterns (cf. Tchalenko, 1970) are developed adjacent to the larger scale late brittle faults.

Late brittle folds are mostly small-scale features of little importance. However, marked changes in the strike in an area immediately north of Route 1 appear to result from a large-scale late brittle dextral folding. This structure produces a distinct kink in the Dover Fault. Centimetre-scale late brittle folds and faults of both dextral and sinistral shear sense are also locally common within the Avalon Zone volcanic rocks, particularly in regions adjacent to major faults (e.g., Terra Nova Lake).

The late brittle structures appear to reflect regional dextral displacements, with a possible component of east-northeast–west-southwest shortening being accommodated by the conjugate fault sets. Alternatively, these structures may be accommodating local stresses generated within the fault block regions between major fault strands. The late brittle fault system is similar to that observed farther north along the coast of Bonavista Bay (Holdsworth, 1991). It does not appear to be extensively developed in the region of the Gander–Avalon zone boundary between Maccles Lake and the northern boundary of the study area.

ACKNOWLEDGMENTS

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A REINTERPRETATION OF THE GEOLOGY OF PARTS OF THE HERMITAGE PENINSULA, SOUTHWESTERN AVALON ZONE, NEWFOUNDLAND

S.J. O'Brien, C.F. O'Driscoll[1] and R.D. Tucker[2]
Newfoundland Mapping Section

ABSTRACT

The late Precambrian geological development of the Hermitage Peninsula (Newfoundland Avalon Zone) has been perceived, until now, to record only two principal tectonic events, viz. 1) the widespread deposition of volcanic and sedimentary rocks (Connaigre Bay Group), and 2) the subsequent emplacement, into this succession, of gabbroic and granitic plutons (Hermitage Bay Complex).

The Connaigre Bay Group, previously defined as comprising four conformable formations, is now viewed to contain two successions of significantly different age and tectonic history, separated by an erosional unconformity. The older of these corresponds to the Tickle Point Formation (sensu *O'Driscoll and Strong, 1979), a submarine and subaerial sequence of corundum-normative, locally garnet-bearing, felsic volcanic rocks. This succession was intruded by a complex of gabbroic and granitic plutonic rocks (Furby's Cove intrusive suite) that includes a bimodal dyke swarm. The plutonic and adjacent country rocks were subsequently unroofed, and served as basement to a volcano-sedimentary cover, represented by the Connaigre Bay Group, as redefined herein. Within the study area, this group includes a basal, dominantly sedimentary unit and an overlying, dominantly mafic volcanic unit,* viz. *the Sam Head and Doughball Point formations, respectively (cf. O'Driscoll and Strong, 1979). A younger series of diorites and gabbros (Grole intrusive suite) crosscuts the older intrusive rocks, the basal Connaigre Bay Group unconformity, and the overlying sedimentary and volcanic cover rocks. Field relationships clearly illustrate the protracted geological history of the region and establish the complex and composite nature of the Newfoundland Avalon Zone in the late Precambrian.*

The late Precambrian stratified rocks host two significant volcanogenic sulphide showings and several base-metal-poor, massive pyrite occurrences. Almost all of these are found within a limited stratigraphic interval at or near the top of the Tickle Point Formation.

INTRODUCTION

PURPOSE AND SCOPE

Mapping and integrated U–Pb geochronological investigations on the Hermitage Peninsula are being carried out by the authors as part of a larger undertaking to understand the nature, relationship and timing of late Precambrian magmatic and mineralizing events in the southwestern Avalon Zone of the Newfoundland Appalachians (cf. Williams, 1979). This work has significant implications for regional correlation of stratigraphic units and tectonic events across the entire Avalon Zone.

Field work to date has focussed on determining both the internal and external relationships of the Hermitage Bay Complex (*sensu* Colman-Sadd *et al.*, 1979) and the Connaigre Bay Group (*sensu* O'Driscoll and Strong, 1979). Sampling of these and other stratified and plutonic units on the Connaigre Peninsula and northern Fortune Bay for geochronological and related isotopic studies has been completed and the main results of this work are to be published at a later date. Presented below is a summary of our new interpretation of the tectonic history of part of the Hermitage Peninsula. This combines the results of previous geological mapping (Greene and O'Driscoll, 1976; O'Driscoll, 1977) with our geochronological results (R. Tucker, unpublished data) and follow-up mapping carried out by S. O'Brien and C. O'Driscoll in June, 1991.

REGIONAL GEOLOGICAL SETTING

The Connaigre Peninsula (Figure 1) embodies the southwestern extremity of the Avalon Zone, the easternmost

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[1] Mineral Deposits Section

[2] Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario

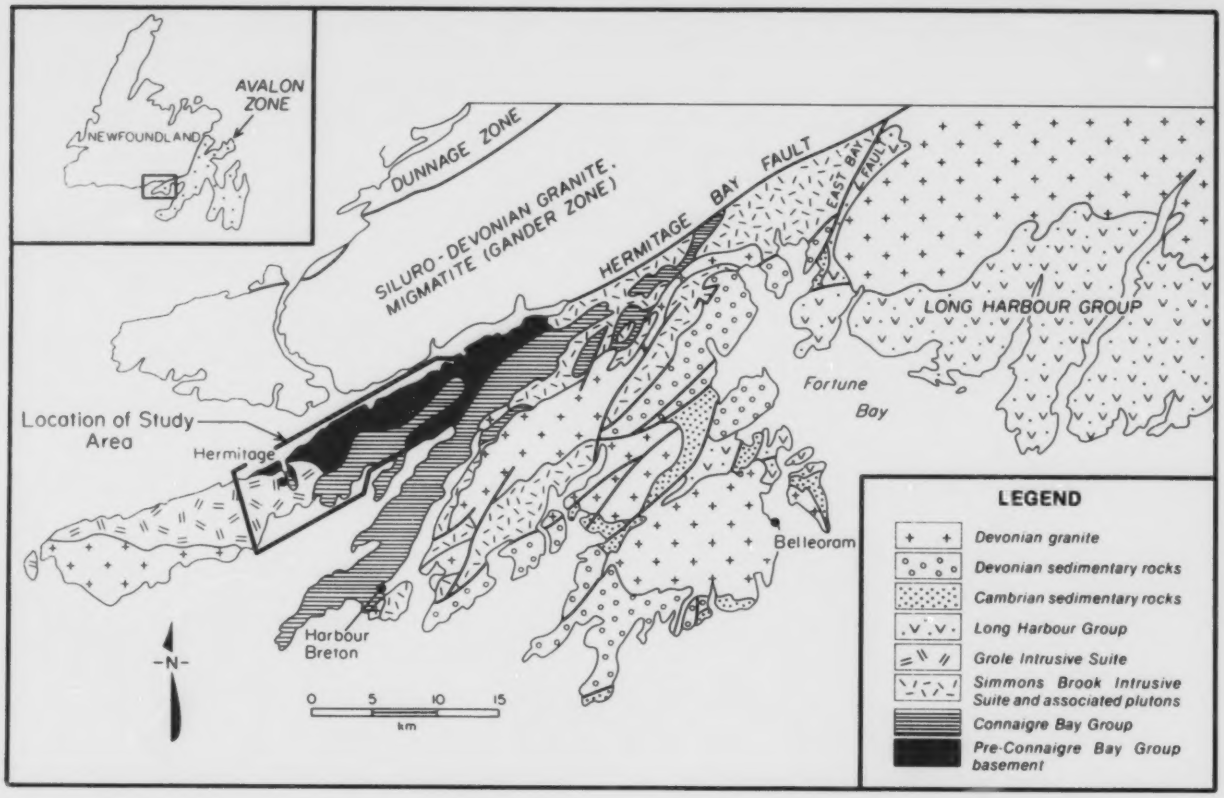
AVALON ZONE
NEWFOUNDLAND
DUNNAGE ZONE
SILURO-DEVONIAN GRANITE, MIGMATITE (GANDER ZONE)
HERMITAGE BAY FAULT
EAST BAY FAULT
LONG HARBOUR GROUP
Fortune Bay
Location of Study Area
Hermitage
Belleoram
Harbour Breton
-N-
0 5 10 15
km
LEGEND
Devonian granite
Devonian sedimentary rocks
Cambrian sedimentary rocks
Long Harbour Group
Grole Intrusive Suite
Simmons Brook Intrusive Suite and associated plutons
Connaigre Bay Group
Pre-Connaigre Bay Group basement

Figure 1. *Generalized geological map of the Avalonian rocks of northwest Fortune Bay, southern Newfoundland.*

of Williams' (1979) tectonostratigraphic divisions of the Newfoundland Appalachians. The Hermitage Bay Fault (Widmer, 1950; Blackwood and O'Driscoll, 1976), a northeast-trending, brittle structure, extends into the offshore under Hermitage Bay (Figure 1), and separates late Precambrian rocks of the Avalon Zone in the southeast from an adjacent lower Paleozoic amphibolite-facies, granite-gneiss terrane in the northwest (Gander and Dunnage zones; Williams, 1979; Colman-Sadd *et al.*, 1990).

The stratified rocks of that part of the Connaigre Peninsula west of the East Bay Fault (Williams, 1971; Figure 1), and more specifically, west of the Simmons Brook Batholith (Williams, 1971) have, in previous studies, been viewed as a single lithostratigraphic entity of undetermined age. Widmer (1950) first mapped these low-grade volcanic and sedimentary rocks and proposed that they be named Connaigre Bay volcanics. Although Widmer (1950) discussed the possibility of the volcanic rocks being Precambrian, he assigned them to the Ordovician on the basis of their similarity to the nearby Long Harbour Series, which White (1939) considered to be of Ordovician age. Greene and O'Driscoll (1976) and O'Driscoll (1977) systematically mapped the Connaigre Bay volcanics and elevated the succession to group status, subdividing it into four formations (see also, O'Driscoll and Strong, 1979; Colman-Sadd *et al.*, 1979). Greene and O'Driscoll (1976) concurred with Widmer's earlier correlation of Long Harbour and Connaigre Bay rocks, but assigned the latter a late Precambrian age, following Williams' (1971) reinterpretation of the age and stratigraphy of the Long Harbour Group.

The Connaigre Bay Group has been viewed as a conformable lithostratigraphic sequence, the base of which is not exposed (cf. O'Driscoll and Strong, 1979). Its stratigraphic subdivisions, as defined by O'Driscoll and Strong (1979), are (in ascending order) the Tickle Point, Great Island, Doughball Point and Down's Point formations (Table 1). Because of nomenclatural duplication, Colman-Sadd *et al.* (1979) subsequently abandoned the name Great Island, replacing it with Sam Head Formation. These units comprise a succession, in excess of 2 km thick, which consists of a lowermost felsic volcanic sequence, overlain by shallow marine tuffaceous sedimentary rocks, which are in turn succeeded by mafic volcanic rocks (O'Driscoll, 1977; O'Driscoll and Strong, 1979). The succession is capped by terrestrial, red to grey clastic rocks.

The Connaigre Bay Group was viewed by all previous workers to be bounded by either faults or intrusive contacts. All intrusive rocks on the Hermitage Peninsula were thought to postdate the deposition of its youngest strata. O'Driscoll and Strong (1979) placed these intrusive rocks into two separate groupings, characterized by distinct ages, petrology and geochemistry. The older group of intrusions were

Table 1. Table of formations for part of the Hermitage Peninsula, comparing present and previous views of the area's geological history

O'Driscoll and Strong, 1979	*This report*
ORDOVICIAN OR EARLIER	**LATE PRECAMBRIAN**
HERMITAGE COMPLEX	GROLE INTRUSIVE SUITE
silicic and mafic dykes	(unseparated)
-----*intrusive contact*-----	-----------*intrusive contact*-----------
Furby's Cove Granite	SIMMONS BROOK INTRUSIVE SUITE
-----*intrusive contact*-----	(unseparated)
Grole Diorite	-----------*intrusive contact*-----------
-----------*intrusive contact*-----------	CONNAIGRE BAY GROUP
PRECAMBRIAN	Down's Point Formation
CONNAIGRE BAY GROUP	-----*conformable contact*-----
Downs Point Formation	Doughball Point Formation
-----*conformable contact*-----	-----*conformable contact*-----
Doughball Point Formation	Sam Head Formation
-----*conformable contact*-----	-----------*unconformity*-----------
Great Island Formation	FURBY'S COVE INTRUSIVE SUITE
-----*conformable contact*-----	Hermitage dykes
Tickle Point Formation	-----*intrusive contact*-----
-----------*base not exposed*-----------	unnamed granite unit
	-----*intrusive contact*-----
	unnamed gabbro unit
	-----------*intrusive contact*-----------
	TICKLE POINT FORMATION
	-----------*base not exposed*-----------

interpreted to be Precambrian in age and comagmatic with Connaigre Bay volcanism; these are the Hermitage Bay Complex (Colman-Sadd *et al.*, 1979; former Hermitage Complex of Greene and O'Driscoll, 1976) and the Straddling Granite (O'Driscoll, 1977; O'Driscoll and Strong, 1979). These plutons were correlated with the Simmons Brook Batholith (Williams, 1971), which intrudes the Connaigre Bay Group east of the Hermitage Peninsula, and lies nonconformably below Devonian strata (Williams, 1971). The younger intrusions locally crosscut lower and mid-Paleozoic strata and postdate the regional deformation of both the Precambrian and Paleozoic rocks. This younger suite consists of the Belleoram, Harbour Breton, Ackley and Pass Island granites (Strong *et al.*, 1974), the last of which is exposed at the southwestern tip of the Hermitage Peninsula. The Pass Island Granite intrudes rocks assigned by previous workers to the Hermitage Bay Complex.

THE PRESENT STUDY

Zircon geochronology by Swinden and Hunt (1991) first challenged previous correlations of the Connaigre Bay Group with other late Precambrian volcanic units of the western Avalon Zone (cf. Colman-Sadd *et al.*, 1990). Subsequent investigation by ourselves (R. Tucker, unpublished data), likewise challenged earlier correlations and further highlighted the need to reassess the nature of the Connaigre Bay Group and its contacts.

Critical, newly discovered and previously known field relationships, supported by geochronological data, demonstrate that rocks of significantly different age and tectonothermal history have, in the past, been erroneously included in a single lithostratigraphic unit, the Connaigre Bay Group. Furthermore, our data demonstrates that plutonic rocks adjacent to the Connaigre Bay Group, which were previously assigned to a single intrusive complex (Hermitage Bay Complex), are divisible into intrusions that either predate or postdate that group's deposition. Finally, additional evidence (R. Tucker, unpublished data) supports the existence of a third separate suite of late Precambrian plutonic rocks (Simmons Brook Batholith of Williams, 1971), which is coeval with the deposition of the Connaigre Bay Group.

REVISIONS AND REDEFINITIONS

On the basis of data summarized below, the Connaigre Bay Group (Table 1) is redefined to exclude the Tickle Point Formation. The newly defined group is herein viewed to consist of (in ascending order) the Sam Head Formation, the Doughball Point Formation, and the Down's Point Formation. The base of the Connaigre Bay Group is now drawn at a newly recognized unconformity at the base of the Sam Head Formation; the stratigraphic top of the group is not observed in the study area. The nature and position of the boundaries between the Sam Head, Doughball Point and Down's Point formations remain largely the same as previously defined (cf. O'Driscoll and Strong, 1979; Colman-Sadd *et al.*, 1979).

The name Hermitage Bay Complex is abandoned, for reasons described below. A new informal name, Furby's Cove intrusive suite, is used to denote those intrusive rocks beneath the sub-Connaigre Bay Group unconformity, specifically: 1) those parts of the former Hermitage Bay Complex (cf. Colman-Sadd *et al.*, 1979) intruded by the former Furby's Cove Granite of O'Driscoll and Strong (1979); 2) the old Furby's Cove Granite; and 3) the bimodal dyke swarm (Hermitage dykes) that intrudes the above rocks. Except for the dykes, the individual components of the suite remain unnamed at present.

The name Simmons Brook intrusive suite is suggested to replace Simmons Brook Batholith (Williams, 1971). This new, informally proposed name refers to the largely pink to grey, medium- to coarse-grained, granitic to tonalitic plutonic rocks of late Precambrian age on the Connaigre Peninsula that are coeval with or intrusive into parts of the Connaigre Bay Group. These plutons are intruded by members of the Grole intrusive suite (see below).

The name Grole Diorite, used by O'Driscoll and Strong (1979) to denote the mafic portion of their Hermitage Complex, is dropped, for reasons outlined below. The new, informally proposed name, Grole intrusive suite, refers to the quartz diorites, diorites, gabbros and related rocks of late Precambrian age that intrude the Connaigre Bay Group, the Simmons Brook intrusive suite and adjacent older units of the Hermitage Peninsula.

TICKLE POINT FORMATION

The oldest rocks on the Hermitage Peninsula (Figure 2) occur within this stratigraphic unit. The formation, the base of which is unexposed, is principally composed of pink to purple and green felsic volcanic rocks that have a buff, and in many cases, rusty colour on weathered surfaces. Massive, and in many places featureless, rhyolite flows containing disseminated pyrite are its most common lithological component; occurring in lesser proportions are flow-banded variants and crystal- and crystal-lithic tuffs of felsic composition. A distinctive petrographic feature is the presence of rare garnet microphenocrysts in some rhyolites (O'Driscoll, 1977). The Tickle Point Formation also contains basalt and andesite flows interstratified with the felsic volcanic rocks.

O'Driscoll (1977) defined the Tickle Point Formation as the basal stratigraphic division of the Connaigre Bay Group, noting that its upper contact with overlying sedimentary rocks was locally marked by conglomerate. In the absence of evidence to the contrary, this upper contact was assumed to be conformable. Our new mapping indicates, however, that the external contacts of the Tickle Point Formation are not only tectonic and intrusive but are also unconformable, rather than conformable. The relationship of the Tickle Point Formation to surrounding rocks (see below), in addition to other regional relationships outlined elsewhere in this account, necessitate the removal of the Tickle Point from the Connaigre Bay Group. Rhyolite of the Tickle Point Formation has yielded zircons that are dated at 682.8 ± 1.6 Ma (Swinden and Hunt, 1991).

FURBY'S COVE INTRUSIVE SUITE

As originally defined by Greene and O'Driscoll (1976), the Hermitage Complex incorporates the gabbroic to granitic rocks that occupy the southwestern half of the Hermitage Peninsula, northeast of the Pass Island Granite. The complex was divided by O'Driscoll (1977) and O'Driscoll and Strong (1979) into the Grole Diorite and Furby's Cove Granite. Colman-Sadd *et al.* (1979) replaced the name Hermitage Complex with Hermitage Bay Complex, retaining the original nomenclature of its subdivisions.

Our recent investigations reiterate earlier observations that the granitic and gabbroic to dioritic phases previously included within the Hermitage Bay Complex are intrusive into the Connaigre Bay Group as defined by O'Driscoll and Strong (1979). However, our work indicates at least three separate intrusive bodies occur within the boundaries of the former Hermitage Bay Complex; two of these are temporally distinct, unrelated plutonic complexes. We employ the new, informal name Furby's Cove intrusive suite to denote those intrusions within the study area that predate the deposition of the Connaigre Bay Group, namely: 1) the granitic body previously referred to as Furby's Cove Granite by O'Driscoll and Strong (1979), 2) the unnamed mafic rocks of the former Hermitage Bay Complex that are intruded by this granite, and 3) a bimodal dyke swarm (Hermitage dykes) sited in 1) and 2) above. All are viewed to be genetically related.

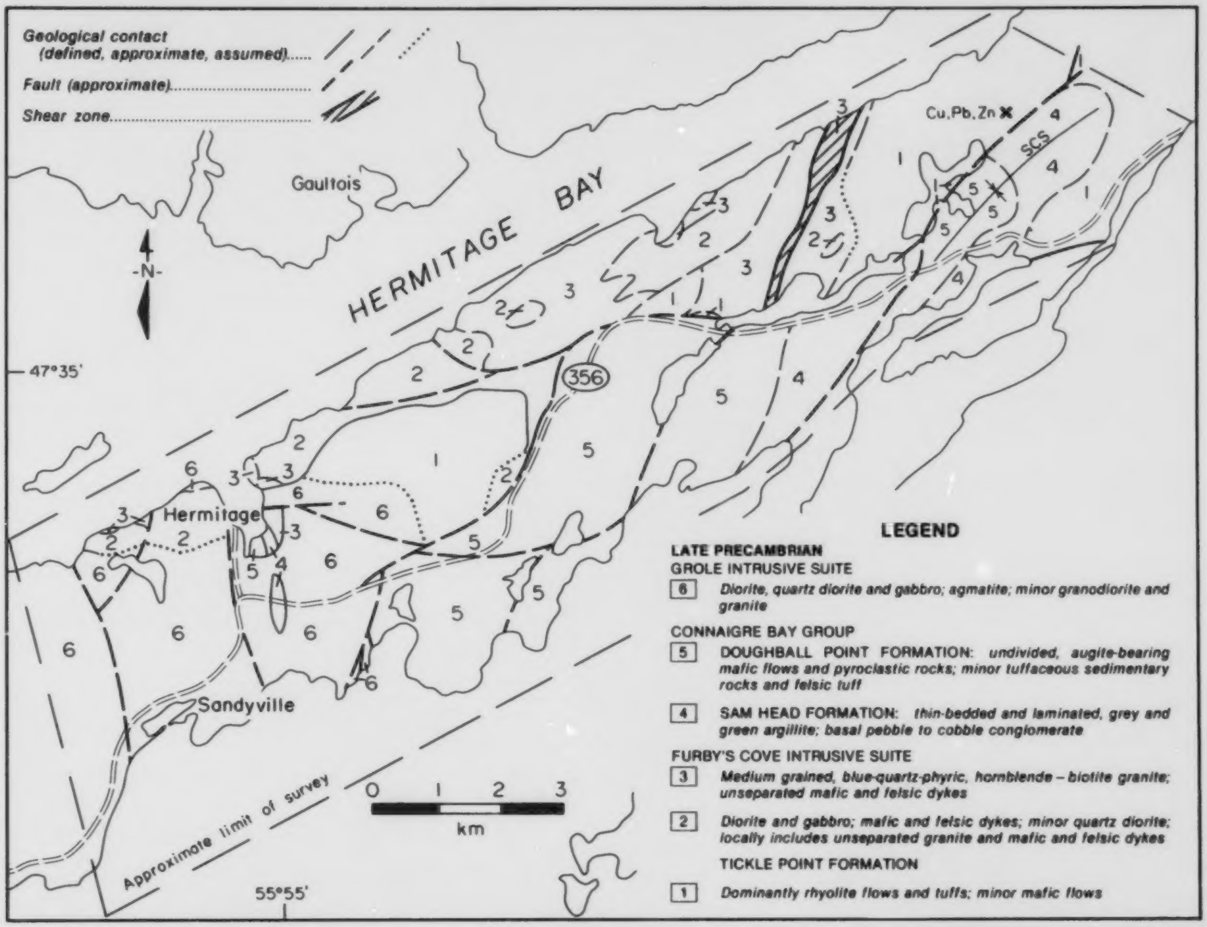
Geological contact (defined, approximate, assumed)
Fault (approximate)
Shear zone
Goultois
HERMITAGE BAY
-N-
47°35'
Hermitage
Sandyville
356
Cu,Pb,Zn
SCS
Approximate limit of survey
0 1 2 3
km
55°55'
LEGEND
LATE PRECAMBRIAN
GROLE INTRUSIVE SUITE
6 Diorite, quartz diorite and gabbro; agmatite; minor granodiorite and granite
CONNAIGRE BAY GROUP
5 DOUGHBALL POINT FORMATION: undivided, augite-bearing mafic flows and pyroclastic rocks; minor tuffaceous sedimentary rocks and felsic tuff
4 SAM HEAD FORMATION: thin-bedded and laminated, grey and green argillite; basal pebble to cobble conglomerate
FURBY'S COVE INTRUSIVE SUITE
3 Medium grained, blue-quartz-phyric, hornblende – biotite granite; unseparated mafic and felsic dykes
2 Diorite and gabbro; mafic and felsic dykes; minor quartz diorite; locally includes unseparated granite and mafic and felsic dykes
TICKLE POINT FORMATION
1 Dominantly rhyolite flows and tuffs; minor mafic flows

Figure 2. *Geological map of part of the Hermitage Peninsula west of Connaigre Bay. X–Frenchman Head showing, SCS–Salmonier Cove syncline.*

UNNAMED GABBRO COMPONENT

The gabbroic phase of the Furby's Cove intrusive suite predates rocks of the Connaigre Bay Group, as presently defined, and postdates the Tickle Point Formation. It was mapped by Greene and O'Driscoll (1976) and O'Driscoll (1977) and included with other mafic intrusions (now found to be younger) in their Hermitage Complex. The rocks are best exposed in coastal sections at and between Furby's Cove and Hermitage. They consist of variably altered, medium-grained, green to dark-grey gabbro to quartz diorite. Mafic minerals consist of subhedral chloritized amphibole and pyroxene. Plagioclase occurs in equal amounts as the mafic minerals and is usually cloudy and corroded. Blue quartz is present where the gabbros and granites are in contact and probably indicates a mixing zone between the two rock types.

UNNAMED GRANITE COMPONENT

The granite phase of the Furby's Cove intrusive suite, first delineated in part by Widmer (1950), was subsequently mapped and named Furby's Cove Granite by O'Driscoll (1977). It consists of a series of narrow sheets and larger bosses of homogeneous, pink to white, equigranular, blue-quartz-bearing granite. It intrudes the older mafic plutonic rocks of the complex and is intimately associated with them. Locally, some field evidence exists for commingling of both granitic and mafic magmas. The granite hosts a major ductile shear zone in the northern part of the study area (Plate 1). There the granite separates the older gabbroic phase of the Furbys Cove intrusive suite from the Tickle Point Formation.

HERMITAGE DYKES

A bimodal dyke swarm, informally referred to here as the Hermitage dykes, forms an integral part of the Furby's Cove intrusive suite. The dykes, which number in the hundreds, occur on a scale of a few centimetres to a decametre in width, have either rectilinear or irregular boundaries, and are diabasic and granitic in composition (Plate 2). The felsite and blue-quartz-phyric dykes are comparable to the fine-grained marginal phases of the granitic component of the

Plate 1. *Mylonitic granite of the Furby's Cove intrusive suite within ductile shear zone, south shore of Hermitage Bay.*

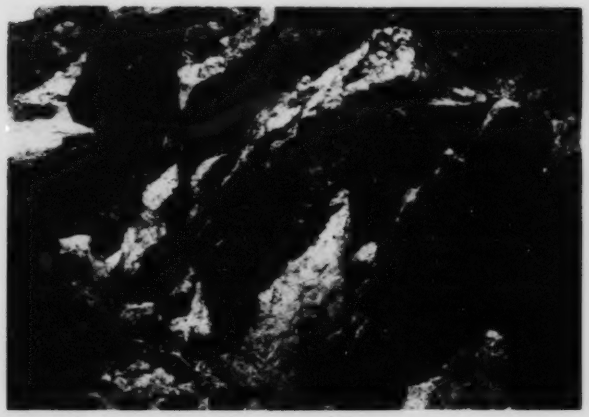

Plate 2. *Tectonized Hermitage dykes, Hermitage Harbour. Field of view approximately 10 m.*

intrusive suite. The dykes occur in both the gabbroic and the granitic components of the suite and, significantly, are sited in a zone parallel to the Hermitage Bay Fault. Similar dykes elsewhere on the Hermitage Peninsula, away from the main dyke swarm, are reported to contain garnet microphenocrysts (O'Driscoll, 1977). Some of the dyke material occurs as detritus in conglomerates nonconformably overlying the Furby's Cove intrusive suite. Other dykes in this zone, however, may be related to younger magmatic events, coeval with deposition of the Doughball Point Formation and emplacement of the Grole intrusive suite.

The external boundaries of the Furby's Cove intrusive suite are either intrusive or unconformable contacts (Figure 2). Each of the three components of the suite intrude the Tickle Point Formation. A newly documented, erosional unconformity locally separates the granite component of the suite from overlying sedimentary and succeeding volcanic rocks. These stratified rocks, dated as late Precambrian (R. Tucker, unpublished data), comprise the Sam Head and Doughball Point formations of O'Driscoll and Strong (1979), respectively. Those parts of the former Hermitage Bay Complex that intrude these two formations are now known to constitute a younger plutonic suite, separate from and intrusive into the Furby's Cove intrusive suite, as defined herein.

CONNAIGRE BAY GROUP

SAM HEAD FORMATION

The name Sam Head Formation was proposed by Colman-Sadd *et al.* (1979) to denote the succession of sedimentary rocks within the Connaigre Bay Group that overlies the Tickle Point Formation. Within the boundaries of the present study area (Figure 2), these rocks occur in the Salmonier Cove syncline (O'Driscoll, 1977) and as large enclaves within diorite in and around the community of Hermitage. In both areas, these sedimentary rocks are conformably overlain by mafic volcanic rocks of the Doughball Point Formation (O'Driscoll, 1977).

The characteristic rock types of the Sam Head Formation, within the study area, are homogeneous, grey and green, laminated argillites and sandstones containing rare limestone lenses (O'Driscoll, 1977). These may be graded, crossbedded, rippled or massive. Polymictic conglomerates are, in places, developed either at or near the base of the formation. Locally, such as in some exposures in the Salmonier Cove syncline, the conglomerates are red and purple.

The Sam Head Formation unconformably overlies both the granitic phase of the Furby's Cove intrusive suite and the Tickle Point Formation. On the southeast shore of Hermitage Harbour, a thin succession of Sam Head Formation is preserved as an enclave in diorite of the Grole intrusive suite. The sedimentary rocks lie in depositional contact with granite of the Furby's Cove intrusive suite, and are overlain by basalt of the Doughball Point Formation. The basal conglomerate of the Sam Head Formation contains rounded to subangular blocks of the directly underlying granite (Plate 3). In addition

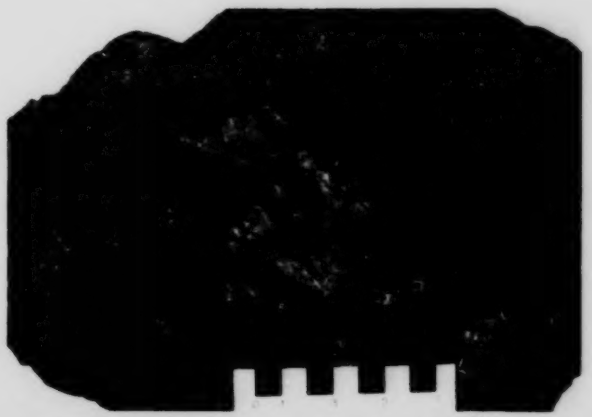

Plate 3. *Basal conglomerate of the Connaigre Bay Group, containing clasts of granite from the underlying Furby's Cove intrusive suite, Hermitage Harbour.*

to granite detritus, the basal conglomerate also includes clasts of diabase and felsite, presumably derived from the bimodal dyke swarm that forms part of the Furby's Cove intrusive suite in the immediate area. The surface of the unconformity has been affected by minor brittle faulting and is the site of intrusion of a late plagioclase-porphyry dyke.

The base of the Sam Head Formation in the Salmonier Cove Syncline is marked by coarse-grained conglomerate, which, in the core of the structure, lies directly upon rhyolite of the Tickle Point Formation. Angular discordance has not been documented at the contact at this locality, because bedding orientations are not developed in the underlying volcanic rocks. The relationships between the Sam Head Formation and granite of the Furby's Cove intrusive suite (which intrudes the Tickle Point Formation), indicate that the Sam Head–Tickle Point depositional contact represents a significant tectonic break.

DOUGHBALL POINT FORMATION

The name Doughball Point Formation (O'Driscoll, 1977; O'Driscoll and Strong, 1979) denotes the predominantly volcanic succession of chiefly basaltic composition within the Connaigre Bay Group that overlies the sedimentary rocks of the Sam Head Formation. O'Driscoll and Strong's (1979) formational definition and boundaries are retained here.

The Doughball Point Formation is characterized by grey and green, mafic flows and pyroclastic rocks but also contains a minor component of felsic volcanic material (O'Driscoll, 1977). Both the massive flows and the pyroclastic rocks, which are dominantly agglomerates, are of basaltic to andesitic composition and commonly pyroxene-phyric.

Mafic volcanic rocks of the Doughball Point Formation conformably overlie the Sam Head Formation in the Salmonier Cove syncline. The contact is well preserved south of the map area at Great Island (O'Driscoll, 1977). Augite-porphyritic mafic dykes that intrude the Sam Head Formation are compositionally similar to the overlying Doughball Point Formation flows and are viewed as feeders to the flows (O'Driscoll, 1977). Mafic hornfels is widely developed in the Doughball Point Formation adjacent to the Grole intrusive suite. Contact migmatites are locally developed at this boundary (Plate 4; O'Driscoll, 1977).

SIMMONS BROOK INTRUSIVE SUITE

The Simmons Brook intrusive suite is a new, informal name, designating the granitic and tonalitic rocks that intrude the Connaigre Bay Group and older sequences, and is overlain nonconformably by Devonian strata (cf. Williams, 1971). The suite lies east of the Hermitage Peninsula study area (Figure 1), but it is interpreted to be coeval with volcanic rocks of the Connaigre Bay Group in that area. In several places, intrusive rocks of mafic composition are spatially associated with felsic phases of the Simmons Brook intrusive suite. Some of these are related to the younger Grole intrusive suite, but others may be comagmatic with the Simmons Brook felsic rocks. Still others may be mafic parts of the Furby's Cove intrusive suite.

Plate 4. *Migmatite developed at the contact of the Grole intrusive suite and the Connaigre Bay Group, roadcut on Route 356 east of Hermitage.*

The various plutonic members of the Simmons Brook intrusive suite have not been defined fully, and the principal batholith comprising this suite (Williams, 1971; Greene and O'Driscoll, 1976) was not studied during the course of the 1991 field investigations. Our present understanding of the regional geological setting of these rocks is based on previous geological mapping (Williams, 1971; Greene and O'Driscoll, 1976), coupled with lithogeochemical and U–Pb geochronological data (O'Driscoll, 1977; R. Tucker, unpublished data).

GROLE INTRUSIVE SUITE

The Grole intrusive suite is an informal name designating the gabbroic, dioritic and related plutonic rocks of the former Hermitage Bay Complex that are intrusive into Sam Head and Doughball Point formations of the Connaigre Bay Group. Its component parts are unseparated at present, and are not identified on Figure 2.

The suite consists of black and dark-green to grey, medium- to coarse-grained gabbro, diorite, quartz diorite and granodiorite. Coarse-grained, black gabbro consists of large crystals of euhedral, largely unaltered, amphibole, pyroxene, biotite and plagioclase. Medium- to fine-grained phases are mineralogically similar. The quartz diorite contains up to 10 percent quartz.

Lithologically, these younger mafic rocks differ from the older Furby's Cove intrusive suite in that they are relatively homogeneous and unaltered, contain significantly fewer dykes, and are associated with the distinctive, coarse-grained, black gabbro phase. Diorite and quartz diorite that together form the chief component of the younger complex are, in many localities, net-veined by granite, which locally coalesces to form small bosses. In at least two localities, the younger diorite displays an incipient, centimetre-scale, gneissic banding. The banding is interpreted to be the result of intrusion into actively deforming zones within and at the margins of the diorite body. The Grole intrusive suite intrudes the Tickle Point Formation, the Furby's Cove intrusive suite, and the stratigraphically higher Sam Head and Doughball Point formations (Greene and O'Driscoll, 1976; Figure 2), as well as the intervening unconformity. The timing of the intrusion with respect to regional structures such as the Salmonier Cove syncline is unknown. However, the Grole intrusive suite crosscuts moderately dipping strata above the unconformity at a high angle, indicating that perhaps some deformation of the Sam Head–Doughball Point cover occurred prior to its intrusion.

STRATIGRAPHIC POSITION OF KNOWN MINERALIZATION

The late Precambrian stratified rocks are host to two significant volcanogenic sulphide showings (Frenchman Head and Winter Hill) and several base-metal-poor, massive pyrite occurrences (Sears, 1990). Almost all the occurrences are found within a limited stratigraphic interval.

The Frenchman Head showing consists of three mineralized outcrop areas that represent a single stratigraphic horizon (Sears, 1990) near the top of the Tickle Point Formation (Figure 2). Mineralization consists mainly of pyrite, sphalerite, chalcopyrite and galena, disseminated within silicified felsic volcanic rocks (Sears and O'Driscoll, 1989). The mineralized felsic rocks are overlain by carbonatized and chloritized mafic volcanic rocks.

The Winter Hill showing, which is located northeast of the study area, is a stratabound deposit that is presently mapped within the lower part of the Sam Head Formation. It consists of a lower stringer zone containing chalcopyrite and pyrite and an upper zone containing disseminated to layered, massive sphalerite, galena, pyrrhotite, pyrite, and chalcopyrite. The true thickness of the mineralized zone is 3 to 9 m (Sears, 1990). Diamond drilling shows that the zone at Winter Hill is underlain chiefly by fragmental felsic and silicified rocks, having lesser amounts of mafic tuffs and flows (Sears, 1990). The drillhole data could possibly indicate that mineralization at Winter Hill is located at the top of the dominantly silicic Tickle Point Formation, below the unconformity at the base of the Sam Head Formation. Sears (1990) stated that the Winter Hill and associated showings in the area represent deposition during a hiatus in volcanism and a change from a felsic-dominated to a mafic-dominated environment. Quartz cobble conglomerates are known to exist in the area and require re-examination to determine if they represent a major break in the stratigraphic section.

Lead isotope ratios of samples taken from the Frenchman Head and Winter Hill showings indicate that both deposits formed during the same mineralizing event and contain lead that came from the same source (S. Swinden, personal communication, 1991). It is significant that one of the few-known equivalents of the Tickle Point Formation in the Avalon Zone elsewhere in the Canadian Appalachians is host to the former Stirling base-metal mine in Nova Scotia. Rocks of the Stirling belt have been dated by U–Pb (zircon) method at 674 ± 1 Ma (Barr *et al.*, 1991). These rocks are close to the age of the Tickle Point Formation of (682.8 ± 1.6; Swinden and Hunt, 1991), and like that formation, are significantly older than the more typical 630 to 600 Ma and 580 to 550 Ma Avalonian sequences (cf. Bevier and Barr, 1990; O'Brien *et al.*, 1990).

SUMMARY AND DISCUSSION

The stratigraphic and intrusive relationships outlined above have necessitated our redefinition of several of the major lithostratigraphic and lithodemic units of the Hermitage Peninsula.

The Connaigre Bay Group, as previously defined by O'Driscoll and Strong (1979), is now known to contain two successions of notably different age and tectonic history, separated by an erosional unconformity. The older of these is an extensive sequence of corundum-normative, locally garnet-bearing, felsic volcanic rocks, assigned by O'Driscoll and Strong (1979) to the Tickle Point Formation. The original name and rank of this unit has been retained; however, it has been excluded from the redefined Connaigre Bay Group.

The submarine and subaerial volcanic rocks of the Tickle Point Formation were intruded by gabbroic to granitic plutonic and hypabyssal rocks of the Furby's Cove intrusive suite prior to the deposition of a volcano-sedimentary cover, the redefined Connaigre Bay Group. A genetic relationship may exist between some or all parts of the Furby's Cove intrusive suite and the Tickle Point Formation; the former possibly represents the magmatic roots of an arc recorded by the Tickle Point volcanic rocks. The magmatic equivalents of the Connaigre Bay cover sequence are thought to be the granitic to tonalitic plutonic rocks of the Simmons Brook intrusive suite.

The bimodal, Hermitage dykes represent a major period of extension within the pre-Connaigre Bay Group basement that is unrelated to subsequent Connaigre Bay Group volcanism. The positioning of the mid-Paleozoic, Hermitage Bay Fault adjacent to this dyke swarm is significant, and may indicate that the zone-bounding fault is located near the position of an older, tectonically weakened zone within the pre-Connaigre Bay Group basement.

The relationship of the ductile shear zone within the granitic phase of the Furby's Cove intrusive suite to rocks above the sub-Connaigre Bay Group unconformity is, as yet, unknown. Ductile fabrics are not developed in any of the Connaigre Bay Group rocks exposed along the projected strike of the north-northeast-trending zone of mylonite. The shear zone has been mapped southward from the coast of Hermitage Bay to a point near the contact of the Tickle Point Formation with the Connaigre Bay Group. The possibility remains that ductile fabrics within the Furby's Cove intrusive suite predate deposition of these younger Precambrian rocks; this hypothesis is to be tested in future field investigations.

In this part of the Hermitage Peninsula, the redefined Connaigre Bay Group includes a basal, dominantly sedimentary unit (Sam Head Formation) and an overlying, dominantly mafic volcanic unit (Doughball Point Formation). The mafic rocks pass conformably upward into red and grey terrestrial clastic sedimentary rocks of the Down's Point Formation (Table 1) elsewhere on the Hermitage Peninsula (O'Driscoll, 1977). Lithologically, this sequence is like many of the late Precambrian volcano-sedimentary assemblages elsewhere in the Newfoundland Avalon Zone, which were deposited at various times between approximately 630 Ma and the end of the Precambrian (Krogh *et al.*, 1988; O'Brien *et al.*, 1990). On the basis of field relationships alone, however, it is not possible to establish with which of these Avalonian volcano-sedimentary groups, if any, the Connaigre Bay Group can be correlated.

In summary, the earliest stages of the late Precambrian geological evolution of the Hermitage Peninsula is now known to involve the formation, extension, uplift and unroofing of a complex arc-like system. These events were followed by deposition of a volcano-sedimentary cover and intrusion of coeval, batholith-scale granitic to tonalitic plutons. Various elements of the basement, as well as the cover and its magmatic roots, were subsequently intruded by significantly younger late Precambrian gabbroic and dioritic plutons. Documentation of the field relationships that illustrate this protracted geological history establishes conclusively the complex and composite nature of the Newfoundland Avalon Zone in the late Precambrian.

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GEOLOGY OF THE NORTHWESTERN PART OF THE GLOVERTOWN MAP AREA (NTS 2D/9) AND THE NORTHEASTERN PART OF THE DEAD WOLF POND MAP AREA (NTS 2D/10)

P. O'Neill
Newfoundland Mapping Section

ABSTRACT

The study area is located within the northeast Gander Zone and is underlain by the Gander Group and a varied suite of syn- and posttectonic granites. The Gander Group consists of interbedded psammite, semipelite and pelite metamorphosed to greenschist and amphibolite facies. Much of the north-central part of the map area is underlain by the posttectonic Siluro-Devonian Gander Lake Granite, a predominantly massive, K-feldspar megacrystic and biotite-rich granite. A static thermal aureole characterized by cordierite and andalusite hornfels occurs peripheral to the granite. The northeastern edge of the Middle Ridge Granite outcrops in the west of the map area and its aureole is characterized by schistose andalusite-bearing rocks.

Southeast of Gambo Pond, the Gander Group is metamorphosed at sillimanite grade and is intruded by a varied suite of granitoid rocks. The latter are typically foliated but the strain is heterogeneous and the fabric ranges from weak to mylonitic. The Hare Bay Gneiss outcrops in a narrow band adjacent to the Gander—Avalon zone boundary.

The Wing Pond Shear Zone, north of Gambo Pond, principally affects metasedimentary rocks of the Gander Group, but on Mint Brook it also deforms a diverse suite of igneous rocks.

INTRODUCTION

During the 1991 field season, the northwestern part of NTS map area 2D/9 (Glovertown; O'Brien *et al.*, 1991) and the northeastern part of NTS map area 2D/10 (Dead Wolf Pond; Blackwood *et al.*, 1991) were mapped at a 1:50 000 scale (Figures 1 and 2). The work represents a continuation of Gander Zone mapping initiated in 1986 in the Weir's Pond area (O'Neill, 1987).

The area south and east of Gambo Pond is accessible from the Trans-Canada Highway (TCH, Route 1), as well as by logging roads and by boat from Maccles Lake and Gambo Pond. The remainder of the area can be reached from Gambo and North ponds and via a system of old logging roads. All-terrain-vehicles were used extensively in the westernmost part of the map area.

The linear nature of Gambo Pond is the most striking physiographical feature in the area. The pond, which is 1 to 2 km wide and approximately 28 km long, is bounded by steep slopes along most of its length, and in the south, near Triton and Riverhead brooks, there are 30- to 60-m-high cliffs. The prominent lineament defined by Gambo Pond appears to continue west-southwest along the valley of Triton Brook. Although no fault breccias were mapped on or near this lineament, it is inferred to mark the locus of a fault. North of Gambo Pond, the country is gently undulating, whereas south and east of the pond, particularly in the vicinity of Gull Pond, the topography is more rugged and reflects, at least partly, the varied geology. Although geologically very significant, the Dover Fault shows no topographic expression. Drainage is mostly east—west and dominated by Triton and Riverhead brooks, which empty into the western end of Gambo Pond. To the north, Mint Brook is the principal waterway draining Mint and North ponds.

Exposure is best north of Maccles Lake and particularly in the area around Gull Pond, where there was a major forest fire in the early 1980's. Despite the steep slopes bordering Gambo Pond, exposure along the north and west shores, especially in the area underlain by the Gander Lake Granite, is poor. The best exposures of the lower-grade part of the Jonathan's Pond Formation occur on Riverhead Brook.

PREVIOUS WORK

The area was surveyed by Jenness (1963) as part of a regional Geological Survey of Canada mapping project. Results of this work were published at four miles to the inch.

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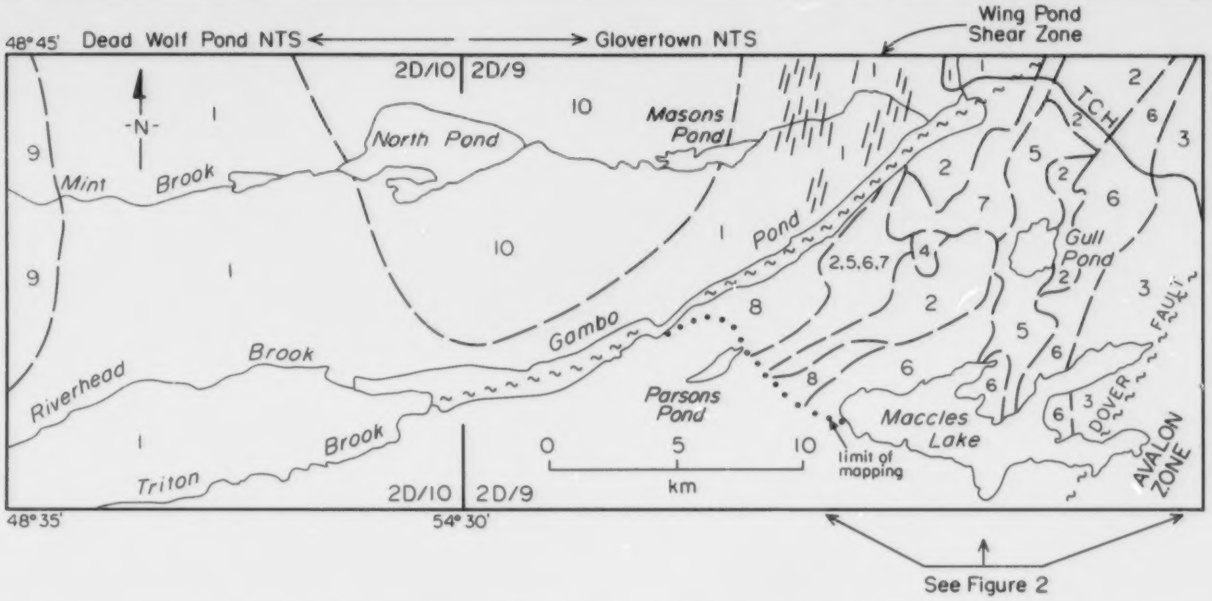

Figure 1. *Simplified geology of the northeast corner of Dead Wolf Pond (NTS 2D/10) map area and the northwest corner of Glovertown (NTS 2D/9) map area.*

The area between Gambo Pond and Gull Pond was interpreted to be underlain by an outlier of the Ackley batholith *(sensu lato)*. Mica quartz paragneiss and mica schist farther east were attributed to the lower unit of the Gander Lake Group. West of Gambo Pond, gneiss, schist and lower-grade equivalents were included in the same unit, and the Gander Lake Granite, which intrudes them, was considered to be another outlier of the Ackley batholith.

Strong *et al.* (1974) carried out a regional lithogeochemical survey of the Gander Zone. They proposed the name Gander Lake pluton to denote a coarse-grained to megacrystic granite outcropping on the Trans-Canada Highway and the shores of Gander Lake. Based on reconnaissance mapping near Maccles Lake, they gave the name Freshwater Bay pluton to exposures of megacrystic granite north of Maccles Lake. Part of the area was also mapped by Blackwood (1976), during a study of the relationships between the Gander and Avalon zones; he renamed the Freshwater Bay pluton as the Maccles Lake granite.

The northwest portion of the Dead Wolf Pond map area, underlain by the Middle Ridge Granite and the Gander Group, was mapped by Blackwood and Green (1982a) and adjoins the present project area.

REGIONAL RELATIONSHIPS

The area lies within the Gander Lake Subzone of the Gander Zone (Williams *et al.*, 1988). The Ordovician and/or older Gander Group (Kennedy and McGonigal, 1972) underlies approximately 30 percent of the map area. These rocks have been traced northeastward to the Carmanville (NTS 2E/8) map area (Currie and Pajari, 1977) and southwestward, for approximately 80 km, to the Mount Sylvester (NTS 2D/3) map area (Dickson, 1987). To the east, the Gander Group passes gradationally into the Square Pond and Hare Bay gneisses and to the west, the group is juxtaposed with the Gander River Complex and Davidsville Group of the Exploits Zone (Blackwood, 1978).

Much of the central part of the map area is underlain by the Gander Lake Granite, a large posttectonic megacrystic intrusion. The northeastern edge of the Middle Range Granite (Strong *et al.*, 1974) was mapped in the western portion of the area. This granite forms a northeast-trending, late kinematic, linear intrusion and is divisible into several phases (Blackwood and Green, 1982a).

The Gander Group is separated into two belts by the Wing Pond shear zone, a major structural feature that extends northward from Gambo Pond to the region north of Wing Pond (O'Neill, 1991).

GENERAL GEOLOGY

JONATHAN'S POND FORMATION (UNITS 1 AND 2)

Approximately 50 percent of the area is underlain by metasedimentary rocks belonging to the Jonathan's Pond Formation of the Gander Group. This formation consists of psammite, semipelite and pelite and, in this area, is separated into two units of contrasting metamorphic grade.

Unit 1 is metamorphosed at chlorite grade, except close to the Gander Lake and Middle Ridge granites, where the rocks are hornfels and fine- to medium-grained schists, respectively. Psammite is weathered light-grey and is typically

LEGEND

DEVONIAN

GANDER LAKE GRANITE

10 *Massive, coarse-grained, K-feldspsar megacrystic, biotite granite*

SILURO-DEVONIAN

MIDDLE RIDGE GRANITE

9 *Massive, coarse-grained, feldspar porphritic, muscovite–biotite granodiorite; minor aplite and pegmatite*

MACCLES LAKE GRANITE

8 *Massive, coarse-grained, K-feldspar megacrystic, biotite granite*

SILURIAN(?)

7 *Foliated, fine- to medium-grained, muscovite–biotite, garnetiferous leucogranite; locally contains unseparated metasedimentary rocks (< 30 percent of exposure) and megacrystic granite*

6 *Foliated, K-feldspar megacrystic, biotite granite: locally intruded by minor leucogranite (Unit 7); locally contains minor metasedimentary rocks*

5 *Foliated, fine-, medium- and coarse-grained leucogranite; coarse-grained to K-feldspar megacrystic, biotite granite; contains minor unseparated metasedimentary rocks*

SILURIAN AND OLDER

4 *Foliated, predominantly fine- to medium-grained leucogranite and metasedimentary rock occurring in equal proportions*

EARLY ORDOVICIAN OR OLDER

HARE BAY GNEISS

3 *Unseparated paragneiss and orthogneiss; minor amphibolite*

GANDER GROUP

JONATHAN'S POND FORMATION (Units 1 and 2)

2 *Predominantly coarse-grained, sillimanite-bearing, pelitic and semipelitic schists; minor amounts of unseparated, foliated, fine- to coarse-grained, and megacrystic granite*

1 *Chlorite-grade psammite, semipelite and pelite; unseparated hornfels and schist in the aureoles of the Gander Lake and Middle Ridge granites; minor unseparated quartzite and metabasic rocks*

SYMBOLS

Geological contact	— — — — —
Foliation, second deformation (inclined)	⇀
Foliation, third deformation (inclined, vertical)	⇀ ⇹
Foliation, high strain (inclined)	⌃
Fault	~ ~ ~ ~ ~
High-strain zone	≡≡≡
Shear sense as implied by kinematic indicators in granitoid	⇆

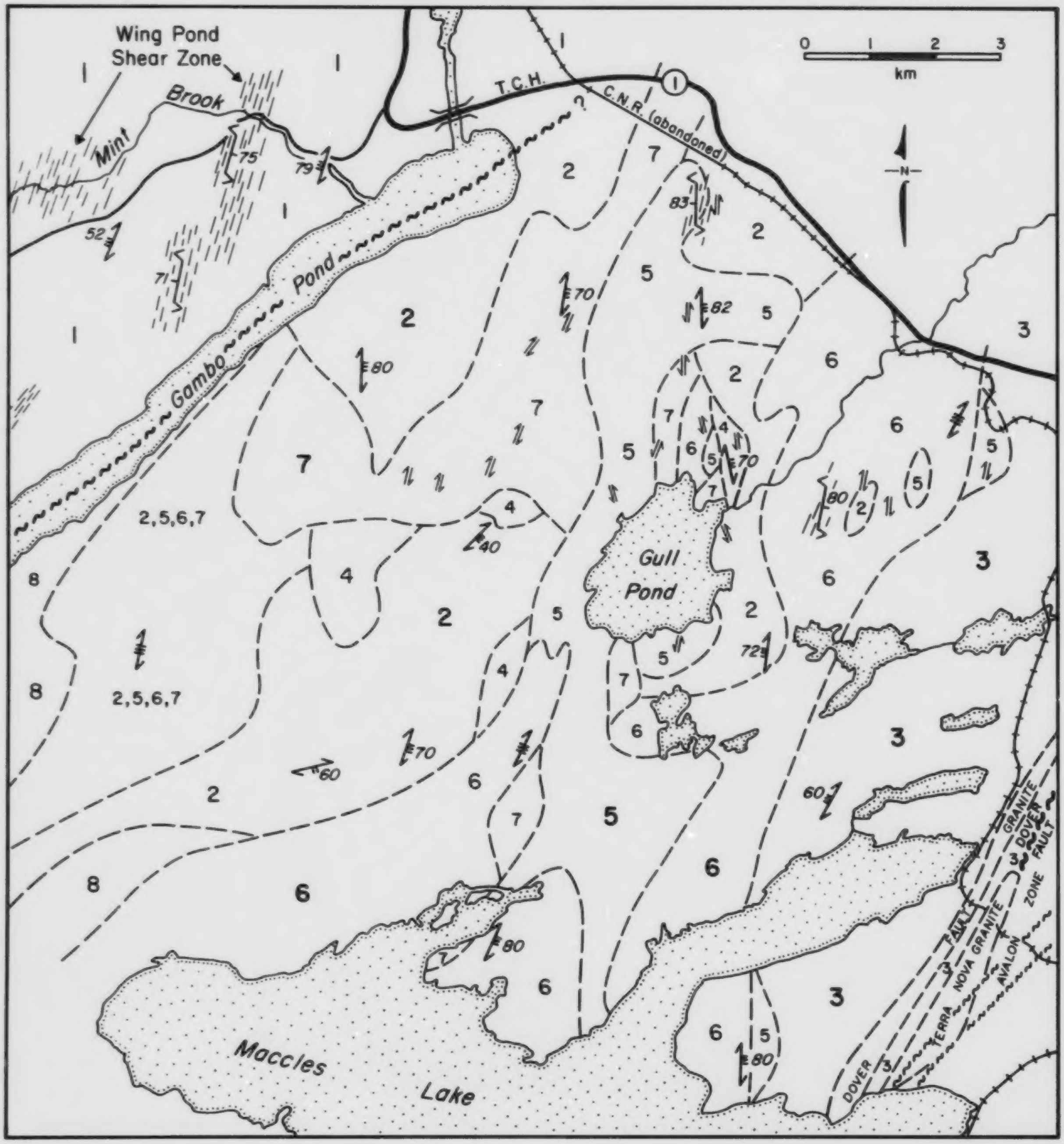
Wing Pond
Shear Zone
Mint
Brook
T.C.H.
C.N.R. (abandoned)
1
0
1
2
3
km
N
Gambo
Pond
Gull
Pond
Maccles
Lake
2,5,6,7
DOVER
FAULT
GRANITE
TERRA NOVA GRANITE
AVALON
ZONE
75
79
52
71
83
70
82
80
40
72
60
80
80
80
70

Figure 2. *Detailed geological map of the Gambo Pond–Gull Pond–Maccles Lake area in the Glovertown (NTS 2D/9) map area.*

grey to greyish-green on fresh surfaces. Semipelite and pelite are weathered light-green and are dark greyish-green on fresh surfaces. Quartzitic beds and mafic bodies form a minor component of the formation. Sedimentary features, other than bedding, are generally masked by the pervasive tectonic foliation and metamorphic recrystallization.

The thermal aureole of the Gander Lake Granite is approximately 2 km wide. The hornfels weathers greyish-white and is pitted; fresh surfaces are dark-blue to black and, in the higher grade areas, purplish, because of the biotite content. Fine- to medium-grained andalusite was noted in several localities. Within 2 to 3 km of the Middle Ridge Granite, the rocks of the Jonathan's Pond Formation are fine- to medium-grained and schistose.

Unit 2 includes rock of the Jonathan's Pond Formation east of Gambo Pond, which are generally coarse-grained

LEGEND

SILURO-DEVONIAN

MACCLES LAKE GRANITE

8 *Massive, coarse-grained, K-feldspar megacrystic, biotite granite*

SILURIAN(?)

7 *Foliated, fine- to medium-grained, muscovite–biotite, garnetiferous leucogranite; locally contains unseparated metasedimentary rocks (< 30 percent of exposure) and megacrystic granite*

6 *Foliated, K-feldspar megacrystic, biotite granite: locally intruded by minor leucogranite (Unit 7); locally contains minor metasedimentary rocks*

5 *Foliated, fine-, medium- and coarse-grained leucogranite; coarse-grained to K-feldspar megacrystic, biotite granite; contains minor unseparated metasedimentary rocks*

SILURIAN AND OLDER

4 *Foliated, predominantly fine- to medium-grained leucogranite and metasedimentary rock occurring in equal proportions*

EARLY ORDOVICIAN OR OLDER

HARE BAY GNEISS

3 *Unseparated paragneiss and orthogneiss; minor amphibolite*

GANDER GROUP

JONATHAN'S POND FORMATION (Units 1 and 2)

2 *Predominantly coarse-grained, sillimanite-bearing, pelitic and semipelitic schists; minor amounts of unseparated, foliated, fine- to coarse-grained, and megacrystic granite*

1 *Chlorite-grade psammite, semipelite and pelite; unseparated hornfels and schist in the aureoles of the Gander Lake and Middle Ridge granites; minor unseparated quartzite and metabasic rocks*

SYMBOLS

Geological contact	— — — —
Foliation, second deformation (inclined)	⟷
Foliation, third deformation (inclined, vertical)	⟷ ⟷
Foliation, high strain (inclined)	⋀—⋁
Fault	~ ~ ~ ~ ~ ~
High-strain zone	≡≡≡≡
Shear sense as implied by kinematic indicators in granitoid	⇆

schists. Dykes of foliated leucogranite and more rarely, megacrystic granite commonly intrude the metasedimentary rocks. The schists occur predominantly in three areas around Gull Pond (Figure 2) and are separated by granitic units. They commonly contain sillimanite, the only metamorphic index mineral observed in the local area. The sillimanite occurs profusely in the exposures in the southern part of the unit, but was not noted in hand specimen farther north, adjacent to Gambo Pond. Generally, it is more profuse eastward in the pelitic and semipelitic rocks. Sillimanite also occurs north of Gull Pond in metasedimentary rocks, which form a minor component in units dominated by leucocratic and megacrystic granite.

HARE BAY GNEISS (UNIT 3)

The Hare Bay Gneiss occurs in a 2- to 3-km-wide band west of the Dover Fault, which forms the Gander–Avalon zone boundary. In the western half of Unit 3, the gneisses are typically coarse grained and of semipelitic composition. Farther east, they are finer grained, tend to be less gneissic in appearance and are characterized by a straight, higher strain foliation.

FOLIATED GRANITES AND METASEDIMENTARY ROCKS (UNITS 4 TO 7)

Unit 4 consists of equal proportions of foliated granite and metasedimentary rocks. Fine- to medium-grained leucogranite is the predominant granitic rock, and it typically contains a foliation that varies widely in intensity. The metasedimentary rocks are coarse-grained schists similar to those of Unit 2 and they locally contain sillimanite.

Unit 5 comprises fine-, medium- and locally coarse-grained leucogranites, and coarse-grained to megacrystic biotite granites. They occur in an irregular-shaped outcrop area that extends north and south from the west side of Gull Pond. Metasedimentary rock containing coarse-grained biotite and muscovite occurs in minor amounts, generally as xenoliths in the granite; no leucosome was noted. The megacrystic granite is invariably cut by the leucogranite, which locally encloses sharply defined blocks of megacrystic biotite granite, containing some tourmaline.

Near the southwest corner of Gull Pond, a white-weathering, medium- to coarse-grained leucogranite exhibits a flow foliation defined by euhedral prisms of feldspar, 0.5 to 1.5 cm across. The foliation is parallel to the regional deformational fabric. One of several subordinate granitic phases in this exposure contains fine-grained tourmaline, which forms up to 20 percent of the rock. Pegmatite veins are common and cut all granite types.

Unit 6 consists of foliated megacrystic granite and occurs principally in two areas, one east of Gull Pond and the other north of Maccles Lake. Feldspar forms 50 to 60 percent of the rock, including abundant megacrysts up to 3 cm long. Anhedral quartz grains, about 1 cm across, form 20 to 30 percent of the granite, and biotite content varies from approximately 5 to 15 percent. Muscovite occurs in only trace amounts.

Metasedimentary material is included in granite of Unit 6 in some places, and it is generally coarse-grained and psammitic to semipelitic in composition. Sillimanite occurs in these rocks but, perhaps because of the composition, is not common.

Some of the granite in the area east of Gull Pond was previously included in the Lockers Bay Granite, a large elongate body of foliated megacrystic granite, which outcrops immediately to the north of the Glovertown area (Blackwood, 1977). Further detailed mapping and geochronological work is required to substantiate this correlation.

A distinctive, high-strain foliation, mostly defined by ribboned quartz, is present in Unit 6. The intensity of this foliation decreases westward. In areas of lower strain, the granite is not readily distinguishable from the megacrystic Maccles Lake Granite (Unit 8).

Unit 7 is a very well-exposed body of foliated leucogranite, which underlies a 5-km-wide area between Gull Pond and Gambo Pond. The granite, which is typically white-weathering and fine- to medium-grained, contains up to 85 percent quartz and feldspar and approximately 15 percent muscovite. Garnet and biotite, although widespread, occur in small amounts, and the biotite is generally restricted to the medium-grained varieties. Tourmaline is locally abundant in garnetiferous pegmatitic veins, which are ubiquitous within the granite.

Coarse-grained, sillimanite-bearing metasedimentary inclusions form 30 to 40 percent of many leucogranite exposures. One outcrop of a fine- to medium-grained, equigranular granite contains numerous micaceous clots that may contain sillimanite. These centimetre-scale aggregates are elliptical and aligned approximately north–south. They may represent restite material derived from pelitic layers of the Jonathan's Pond Formation. Metasedimentary material in the xenoliths does not show evidence of partial melting but quartzofeldspathic veins are common. Megacrystic granite inclusions form up to 30 percent of leucogranite exposures.

LATE SYN- TO POSTTECTONIC GRANITES (UNITS 8 TO 10)

Maccles Lake Granite (Unit 8)

Massive, homogeneous, coarse-grained to megacrystic granite outcrops between the west end of Maccles Lake and Gambo Pond. This granite, which is best exposed on a brook between Parsons Pond and Gambo Pond, was included in the Maccles Lake granite by Blackwood (1976). The granite is compositionally inhomogeneous but in this area contains only minor amounts of mica and traces of magnetite. Metasedimentary xenoliths are present in a few places. A heterogeneous foliation indicates that the granite is late syn-tectonic.

Middle Ridge Granite (Unit 9)

Exposures of granodiorite along the western edge of the area represent the northeastern termination of the Middle Ridge Granite of Blackwood and Green (1982b). The portion of the pluton mapped in this study corresponds to Unit 10 of these authors. The granodiorite is massive and homogeneous, weathers pink and contains 40 percent (or less) quartz, up to 6 mm across, and up to 60 percent feldspar crystals, 2 cm or less in length. Muscovite forms up to 15 percent of the rock, whereas biotite and tourmaline occur in trace amounts only. The granodiorite is intruded by thin dykes of aplite and pegmatite.

Petrography reveals that most of the feldspar is plagioclase, which is extensively hematized and shows minor sericitization. K-feldspar, which forms about 10 percent of the rock, occurs as slightly hematized crystals up to 3 mm across, and contains inclusions of plagioclase and quartz; it is locally perthitic. All grain boundaries are curved, straight or scalloped. Most of the biotite is chloritized.

Gander Lake Granite (Unit 10)

The Gander Lake Granite underlies the central part of the map area and is very poorly exposed. The granite is massive, and is characterized by abundant K-feldspar megacrysts up to 6 cm in length, in a coarse-grained, quartz–feldspar–biotite matrix. The megacrysts commonly contain small inclusions of biotite. White-weathering feldspars form 60 to 70 percent of the rock and are commonly compositionally zoned. Greyish-blue quartz constitutes approximately 20 percent of the matrix. Outcrops of the granite on the small island on the west side of North Pond contain up to 40 percent modal quartz and are cut by abundant quartz veins. Biotite forms 10 to 15 percent of the granite. Accessory tourmaline is locally concentrated in small patches, 5 to 10 cm across. The contact between the granite and the Gander Group is not exposed in the map area.

METAMORPHISM

Sedimentary rocks of the Gander Group west of Gambo Pond are metamorphosed at chlorite grade. Superimposed on this regional metamorphism are two distinct types of contact metamorphism. Rocks adjacent to the Middle Ridge Granite are schistose and contrast texturally with the andalusite-bearing hornfels that surrounds the Gander Lake Granite.

Between Mason's Pond and the northeastern end of Gambo Pond, the Gander Group and spatially associated igneous rocks are mylonitized within the Wing Pond shear zone (O'Neill, 1991). Gander Group phyllonites within this zone contain sillimanite, kyanite and andalusite.

South of Gambo Pond, pelitic and semipelitic rocks of the Gander Group are coarse grained, schistose and contain sillimanite. The sillimanite commonly forms more than 50 percent of pelitic rocks, and is the sole aluminosilicate mineral noted in this area. Exposures of the Gander Group near the north end of Gambo Pond contain no sillimanite.

DEFORMATION

One principal tectonic foliation is defined throughout much of the Gander Group. This fabric is here designated S_2, so as to conform with fabric description established in the Weir's Pond, Gander Lake and Gambo areas to the north (O'Neill, 1990, 1991). S_2 is a slatey cleavage in pelitic rocks and a solution-seam cleavage in psammites. It generally strikes northeast to north-northeast. East and south of the Gander Lake Granite, the fabric dips moderately to steeply northwest or southeast. Between the Middle Ridge and Gander Lake granites, the foliation dip is moderate to shallow to the northwest, and strikes north-northeast. The foliation is steeper near the contact with the granites.

In places, S_2 is folded by mesoscopic-scale, open to tight F_3 folds that are associated with a crenulation lineation. F_3 fold axes plunge shallowly to the north-northeast or south-southwest.

The most prominent structural feature in Unit 1 of the Jonathan's Pond Formation is the Wing Pond Shear Zone, a high-strain zone that extends at least 40 km northward from the north shore of Gambo Pond to the Weir's Pond map area (O'Neill, 1991). In the study area, the shear zone is best exposed east of Mason's Pond, on Mint Brook (Figure 1). Metasedimentary rocks within the shear zone are medium- to coarse-grained phyllonitic schists, and igneous rocks, ranging in composition from gabbro to granite, have a heterogeneous high-strain foliation. Near the eastern edge of the shear zone in Mint Brook, the regional S_2 fabric of the Gander Group is intensely folded by tight to isoclinal folds of psammite and semipelite. These folds have an axial-planar fabric that parallels the high-strain fabric in the adjacent phyllonites and deformed igneous rocks. This suggests that the high-strain foliation may be correlated with S_3 and that the shear zone is related to the D_3 event. The high-strain fabric is typically steep but the presence of moderate dips in some places implies subsequent open folding. Farther east, near the Trans-Canada Highway, at the north end of Gambo Pond, phyllonitic, chlorite-grade rocks of the shear-zone margin pass eastward into low-strain, low-grade, Gander Group pelite and semipelite.

Schists of the higher grade part of the Jonathan's Pond Formation (Unit 2) southeast of Gambo Pond, typically do not show a well-developed foliation. The originally equigranular nature of the psammitic rocks hindered fabric development, whereas fabrics that did form in pelitic beds have been largely obscured by the subsequent development of sillimanite. These features suggest that metamorphism outlasted deformation.

Southeast of Gambo Pond, a weak- to high-strain foliation is developed in most of the granitic rocks and is parallel to a fabric in adjacent metasedimentary rocks. As the latter is axial planar to F_3 folds of the regional S_2 fabric, the foliation developed in the granitic rocks must be younger than the principal tectonic foliation of the Gander Group. Thus, the regional fabric of the granitic intrusions is designated S_3. The intensity of S_3 rocks ranges from weak in the southwest to mylonitic in the northeast.

Well-foliated, fine- to medium-grained granites commonly exhibit good shear-sense criteria. C–S fabrics, associated shear bands and, more rarely, asymmetrical tails on feldspar megacrysts have been used to define three regional zones with differing senses of shearing. C–S fabrics are best displayed in a several-kilometre-wide area of Unit 7 leucogranite northwest of Gull Pond (Figure 2), where the dextral sense of shear is consistent. Farther east, in a mixed complex of granitic and metasedimentary rocks around Gull Pond, the sense of shear is consistently sinistral. East of Gull Pond, however, the sense of motion implied by C–S fabrics, developed mostly in megacrystic granite, is again consistently dextral.

Weakly deformed pegmatite that commonly cuts the granites and the high-strain foliation, may be used to constrain the upper age limit of the deformation. It will require further detailed structural study, however, to determine the relative timing of dextral and sinistral deformations. More detailed study is also required to establish the timing relationships between the Wing Pond and the Dover Fault shear zones, the latter described elsewhere in this volume (O'Brien and Holdsworth, *this volume*).

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DUNNAGE ZONE GRAPTOLITES: AN EXTENSION OF THE AGE RANGE AND DISTRIBUTION OF CERTAIN ORDOVICIAN FORMATIONS OF THE EXPLOITS SUBZONE

S.H. Williams, B.H. O'Brien[1], S.P. Colman-Sadd[1] and F.H.C. O'Brien[2]
Department of Earth Sciences, Memorial University of Newfoundland
St. John's, Newfoundland, A1B 3X5

ABSTRACT

An analysis of Ordovician graptolites from localities in the Dunnage Zone, central Newfoundland, has resulted in refinement of previous correlation and dating of certain strata, of previously unknown age, in the Exploits Subzone. Graptolites recovered from several new fossiliferous localities in the type area of the Upper Ordovician Point Leamington Formation are indicative of the Dicellograptus complanatus *and* Dicellograptus anceps *zones of the Ashgill. The fauna from the latter zone includes* Paraorthograptus pacificus *(Ruedemann) at one locality, demonstrating that the strata belong to the* P. pacificus *Subzone of the late* D. anceps *Zone. Conodonts from detrital clasts in sediments of the Point Leamington Formation and the Goldson Conglomerate indicate the* Amorphognathus superbus *and* Amorphognathus ordovicicus *zones, suggesting that the units are restricted entirely to the Ordovician.*

A sparse Llanvirn fauna, including members of Holmograptus, Xiphograptus *and an unidentifiable diplograptid, occurs in the Strong Island chert of the Exploits and Wild Bight groups, providing the first date for the upper parts of these units. Graptolitic black shale of the Caradoc Lawrence Harbour Formation, which occurs above these medial Ordovician elements of the Exploits Subzone, is also recognizable in the Baie d'Espoir Group on the south coast of Newfoundland.*

INTRODUCTION

In the Exploits Subzone (Figure 1), fossil-bearing units of marine stratified rocks are known to range in age from Tremadoc to Wenlock. Graptolites from black shale facies, present in units as old as Arenig and as young as Llandovery, are particularly useful for biostratigraphic correlation of these Ordovician and Early Silurian rocks. They are also important in establishing the presence or absence of regional variations in the temporal and spatial relationships of laterally discontinuous, diachronous or condensed lithostratigraphical units, especially where such units recur across major structural features. The data presented in this paper exemplify how graptolite faunas can be utilized to document such phenomena, and ultimately aid in our understanding of the tectonic development of the Dunnage Zone.

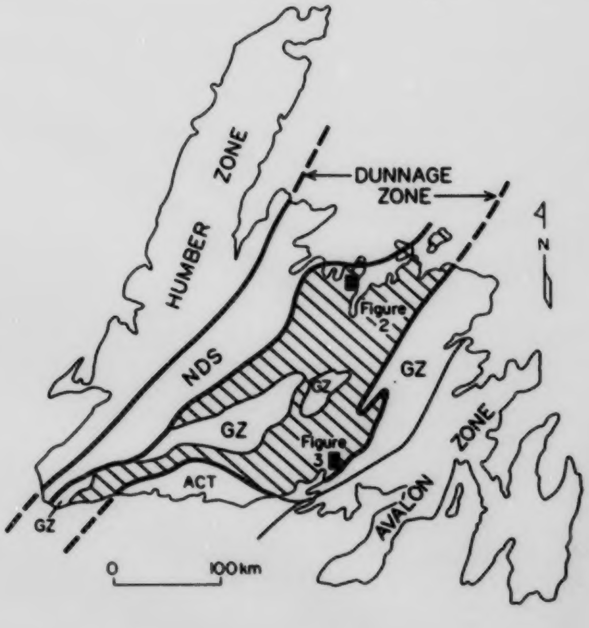

Figure 1. *Locations of the study areas in the Exploits Subzone depicted in Figures 2 and 3. Principal tectonic divisions of the Newfoundland Appalachians are also indicated; Exploits Subzone is patterned, while NDS, GZ and ACT indicate the Notre Dame Subzone, Gander Zone and Avalon Composite Terrane, respectively.*

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[1] Newfoundland Mapping Section, Newfoundland Department of Mines and Energy, Geological Survey Branch

[2] P.O. Box 129, Torbay, Newfoundland A0A 3Z0

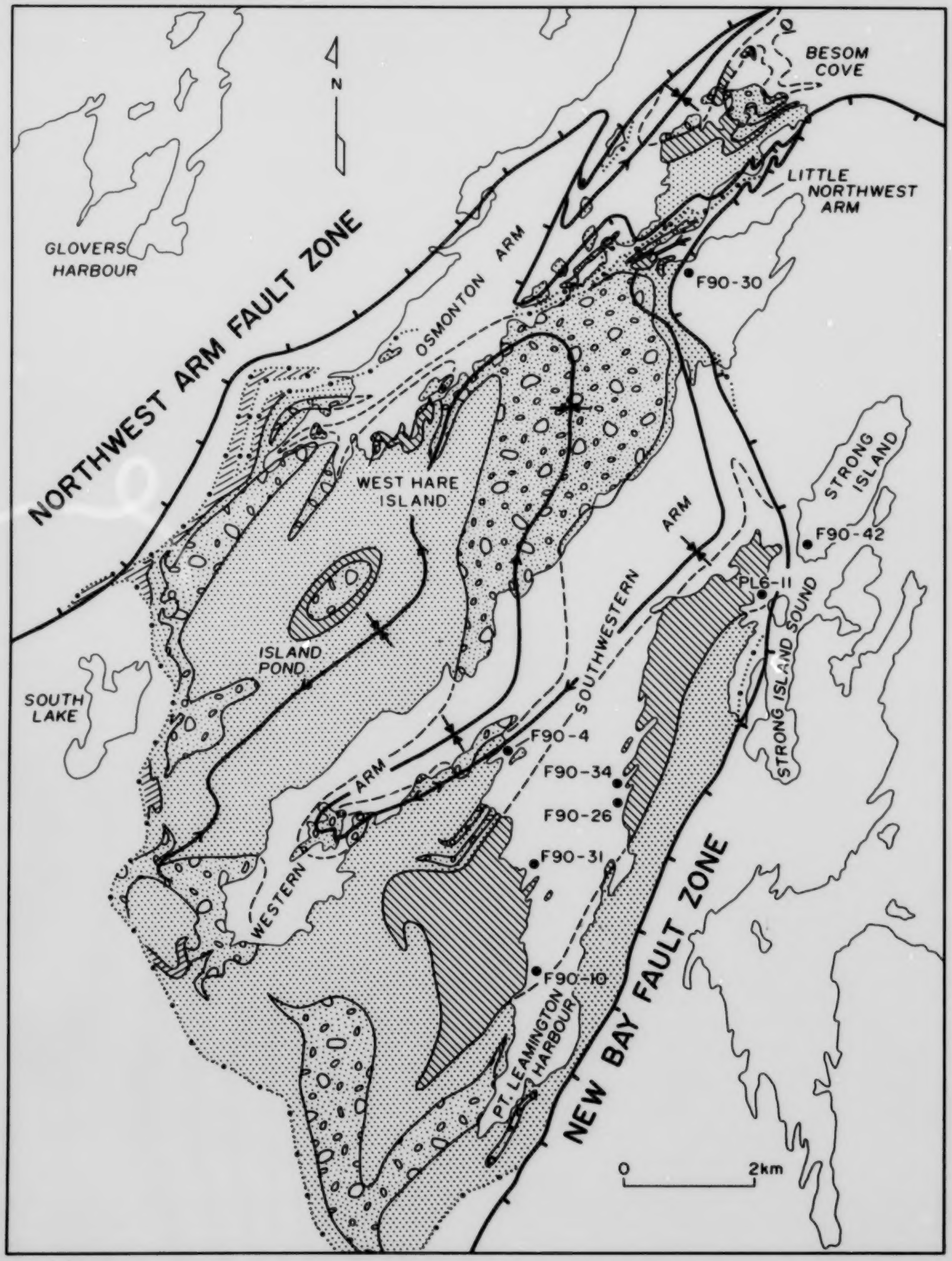
N
BESOM COVE
LITTLE NORTHWEST ARM
GLOVERS HARBOUR
NORTHWEST ARM FAULT ZONE
OSMONTON ARM
F90-30
WEST HARE ISLAND
STRONG ISLAND
F90-42
PL6-11
ISLAND POND
SOUTH LAKE
SOUTHWESTERN ARM
STRONG ISLAND SOUND
F90-4
F90-34
F90-26
WESTERN ARM
F90-31
F90-10
PT. LEAMINGTON HARBOUR
NEW BAY FAULT ZONE
0
2km

Figure 2. *Simplified geological map of the Point Leamington Formation in the Point Leamington basin. Modified, in part, from Geological Survey Branch, Open File maps 90-124 and 91-171 (O'Brien, 1990, 1991b). Graptolite localities F90-26 and F90-34; together with PL6-11 (Williams, 1991b), belong to the* D. complanatus *Zone, and F90-4, F90-10 and F90-31 are indicative of the* D. anceps *Zone of the Point Leamington Formation. Localities F90-30 and F90-42 are from within the Strong Island chert.*

LEGEND

UPPER ORDOVICIAN

POINT LEAMINGTON FORMATION

⁘	Pebbly wacke – polymictic conglomerate lenticles (*P. linearis* Zone to *D. anceps* Zone ?)
▨	Grey shale-dominant lenticles (*D. complanatus* Zone to *D. anceps* Zone ?)
▧	Grey shale-dominant lenticles (*D. clingani* Zone ?)
░	Unseparated turbidite sandstone units

KEY

Lithostratigraphic boundary (defined, assumed)	╱ ╌
Regional fault zone (ticks indicate general dip direction)	┿
Major syncline (plunge direction indicated)	✕
Stratigraphic base/external structural boundary of Point Leamington Formation	⋰
Fossil locality	● F90-30

BIOSTRATIGRAPHY

Ashgill graptolites from the Point Leamington Formation

The definition and graptolite biostratigraphy of the Upper Ordovician Point Leamington Formation of north-central Newfoundland was revised recently by Williams (1991b). The base of this lithostratigraphic unit of siliciclastic flysch was defined on the incoming of the first thick sandstone beds overlying the black shale of the upper part of the Lawrence Harbour Formation (the latter formation being an expansion of the Lawrence Harbour Shale of Helwig, 1969). Where the Lawrence Harbour Formation underlies the Point Leamington Formation, the transition occurs at a level within the *Dicranograptus clingani* or *Pleurograptus linearis* zones, depending upon locality. It was further demonstrated that stratigraphically higher intervals of interbedded shale and sandstone within the formation belong to distinctly different graptolite biozones, and that assemblages indicating the *Dicellograptus complanatus* and *Dicellograptus anceps* zones of Scotland, were present.

More recently, during a Geological Survey Branch multidisciplinary project that included remapping the Point Leamington Formation in its type area, *D. anceps* Zone assemblages have been recovered from eight additional localities and *D. complanatus* Zone faunas from a further two. Most of these sites have yielded graptolites that are also present in the locations reported on by Williams (1991b) and they simply serve to confirm most of his biostratigraphic conclusions. However, such new data, integrated with other unpublished biozonal information, are useful in identifying and mapping certain stratigraphic intervals in the Point Leamington Formation in which fossiliferous dark-grey shale predominates over quartz-wacke in an otherwise monotonous turbidite sandstone succession.

D. complanatus Zone

Faunal assemblages indicative of the *D. complanatus* Zone occur near the base of the largest mappable lenticle of shaly flysch in the Point Leamington Formation. Extending from Strong Island Sound to the vicinity of Southwest Arm (Figure 2), this thinly stratified lenticle of interbedded dark-grey shale and light-grey concretionary sandstone is host to most of the olistostromal deposits in the Point Leamington Formation. Although much smaller than the largest body and to date unyielding of graptolites, possibly contemporaneous lenticles of dark-grey shaly flysch are located in Osmonton Arm near West Hare Island and again near Besom Cove, as well as on the peninsula separating Osmonton and Southwest arms near Island Pond (Figure 2).

Strata of the *D. complanatus* Zone interval provide a younger age limit for several lithological subunits in the lowest 100 to 1000 m of the Point Leamington Formation. Immediately underlying *D. complanatus* Zone strata is a regionally developed sequence marked by distinctive, pyritic, bedding-parallel, worm burrows and other possible trace fossils. The monotonous turbidite sandstone succession below the trace-fossil horizon, but well above the Lawrence Harbour Formation, contains several of the stratigraphically lower lenticles of pebbly and conglomeratic (Goldson) flysch in the Point Leamington basin. Thus, the generalized stratigraphy of the pre-*D. complanatus* portion of the Point Leamington Formation is, in ascending order, thin- to medium-bedded turbidite sandstone, massive pebbly sandstone and polymictic conglomerate, and siltstone–sandstone rhythmites that have conspicuous trace fossils. This succession is capped by the previously discussed olistostrome-bearing shaly flysch lenticles and floored by either the Lawrence Harbour Formation or the oldest (*D. clingani* Zone) grey shale lenticles in the Point Leamington Formation. Characteristic of the marginal regions of the entire Point Leamington basin, and probably in part of *P. linearis* Zone age (Williams, 1991b), it varies considerably in total thickness, although a predictable sequential order of individual subunits is commonly present.

D. anceps Zone

Several taxa in addition to those described by Williams (1991b) have been identified from graptolite collections made in new localities within the *D. anceps* Zone of the Point Leamington Formation. Of these findings, the most important are, from taxonomic and biostratigraphic perspectives, the recovery of mature specimens of *D. anceps* (Nicholson), recognized tentatively by Williams (1991b) only on the presence of one small proximal fragment, and of *Paraorthograptus pacificus* (Ruedemann) from one locality. The latter species is characteristic of the later part of the *D. anceps* Zone (*P. pacificus* Subzone) in southern Scotland (Williams, 1982), and is an invaluable taxon for correlation of Upper Ordovician graptolitic sequences throughout the world.

The lenticle of shaly flysch in the Point Leamington Formation that contains *D. complanatus* Zone assemblages near its base extends stratigraphically upward through the *D. complexus* Subzone and into the *P. pacificus* Subzone of the *D. anceps* Zone. North of the town of Point Leamington, where it displays its maximum thickness (O'Brien, 1990), this lenticle is host to five, stratigraphically unique olistostrome horizons; each of these debris flows is separated by unbroken Point Leamington Formation carrying *in situ D. anceps* Zone assemblages. The youngest dated block of transported material in these olistostromes is Ashgill in age, although Caradoc olistoliths are also present (Williams, 1991b). Slump-folded shale blocks have yielded graptolites belonging to the *P. linearis* Zone, while rip-up clasts of nodular limestone contain conodonts indicative of the *Amorphognathus superbus–Amorphognathus ordovicicus* biozones. Both are consistent with the age of the enclosing or autochthonous strata in this part of the Point Leamington Formation.

In the vicinity of Western Arm, Point Leamington strata that overlie the largest shaly lenticle in the formation contain graptolite assemblages that also place these beds within the *D. anceps* Zone. Faunas characteristic of this biozone were recovered from minor grey-shale intervals in a turbidite sandstone sequence that is interstratified with several small lenticles of polymictic conglomerate and that directly underlie the main body of Goldson Conglomerate in the Point Leamington basin (Figure 2; O'Brien, 1990). Detrital limestone clasts systematically sampled from the uppermost exposed parts of the Goldson Conglomerate have failed to yield conodonts that are any younger than the upper Ashgill, possibly implying that a preserved record of Silurian deposition is not to be found in this region (see also Arnott *et al.*, 1985). The biostratigraphic information discerned from the expanded paleontological database is of fundamental importance in interpreting the lithostratigraphically isolated lenticles of pebbly and conglomeratic flysch within the Point Leamington basin. It successfully demonstrates that these are not contemporaneous tongues emanating from a single, alternately regressive and transgressive, sheet deposit of Goldson Conglomerate but, instead, that such lenticles generally become thicker and more laterally continuous with time. Conglomeratic infilling of the basin, which occurred throughout much of Ashgill time, overlapped most of the period of accumulation of the host greywacke succession.

Arenig–Llanvirn Graptolites from the Strong Island Chert

The Strong Island chert is a regionally widespread unit without formal rank that occurs in both the Exploits and Wild Bight groups (O'Brien, 1990). It overlies, with primary depositional contact, volcanic or sedimentary formations that comprise, in some though not all locations, the upper parts of these undated rock groups (O'Brien, 1991a; Dec *et al.*, *this volume*). The Strong Island chert conformably underlies the basal division of grey-mottled chert in the Lawrence Harbour Formation (Caradoc) and, together, these two units represent the most laterally extensive yet anomalously thin sedimentary packet in the Exploits Subzone. However, in some rare localities, the Lawrence Harbour Formation rests directly on Exploits or Wild Bight formations, which elsewhere form the local substrate of the Strong Island chert.

Recently discovered graptolite occurrences in the Strong Island chert, approximately five kilometres distant from each other along strike, were revisited and recollected in 1991. Fossiliferous intervals located in Little Northwest Arm and in the type area on Strong Island itself are thin, dark-grey, silty laminae occurring within a thin-bedded sequence of grey siltstone and sandstone, most of which are too pale and coarse for graptolite preservation. Within these laminae, most of the carbonaceous material is of uncertain origin; some is, however, composed demonstrably of graptolite fragments, and a few of these are sufficiently well preserved to permit tentative identification.

The majority of fragments represent dendroids (benthic graptolites) of little, if any, biostratigraphic use. Both localities yield a variety of forms, including specimens probably assignable to the genera *Acanthograptus*, *Dendrograptus* and *Dictyonema*. The extremely fragmentary nature of the material does not allow more positive identification; the first two genera range from Cambrian to at least Silurian, while *Dictyonema* is also known from the Carboniferous (Bulman, 1970). The only significance of these forms is paleoenvironmental, as benthic graptolites are apparently restricted to relatively shallow, near-shore conditions. The preservation of their fragile, organic skeletons in a hemipelagic chert and epiclastic turbidite sequence thus indicates relatively limited post-mortem transport of the rhabdosomes from their original setting and rapid burial in a deep oceanic, reducing environment, possibly located on the flanks of an emergent volcanic ridge.

The graptoloid fauna of the Strong Island locality is dominated by fragments of a slender, two-stiped sinograptid, apparently belonging to the species *Holmograptus lentus* (Törnquist). This species is a senior synonym of *H. callotheca* (Bulman) (Jaanusson, 1960; Skevington, 1965); the systematics of this and related species were discussed in full by Archer and Skevington (1973). The original description (Törnquist, 1911) was based on pyritized moulds from the upper Didymograptus Shale of Scania, Sweden. Bulman's (1932, 1936) description of this species was based on three-dimensional, isolated material from the Holen Limestone (variously known as the 'gra Vaginatumkalk', 'Lower Orthoceras Limestone' and 'Asaphus Limestone') of Öland, Sweden, as were most subsequent descriptions (see Skevington, 1965). The Strong Island specimens are, however, very similar in terms of their overall stipe width and form, sicular dimensions and folded dorsal margin. The Holen Limestone is now generally correlated with the top *D. hirundo* or *D. artus* Zone of the United Kingdom (Jaanusson, 1982), suggesting a late Arenig or early Llanvirn age. A detailed stratigraphic discussion of these intervals was given by Skevington (1965), who concluded that the interval yielding *H. lentus* was equivalent to the Llanvirn *D. artus* (ex '*D. bifidus*') Zone of Wales. *H. callotheca* was also described by Bell (1960) from the D_{a3} interval of the Victorian slatebelts in the Lachlan foldbelt of southeast Australia, where a graptolitic flysch also directly overlies an oceanic substrate of mafic volcanic rocks (Gray and Willman, 1991).

The specimens from Strong Island are distinct from other earlier members of *Holmograptus* described from the Cow Head Group of western Newfoundland by Williams and Stevens (1988). Three specimens of a biserial graptoloid were also recovered, but these are unidentifiable even to generic level. The presence of diplograptids does, however, confirm an age of latest Arenig or later.

The graptoloid fauna recovered from Little Northwest Arm is even sparser, consisting of only five specimens sufficiently complete to permit identification. Of these, three are poorly preserved examples of *Holmograptus*. They are similar to those of Strong Island in terms of overall rhabdosome form and width, but lack the folded dorsal margin. This morphological difference may, however, be merely a reflection of poor preservation, and they are thus here referred to *Holmograptus* sp. cf. *H. lentus* (Törnquist). The remaining two graptoloid specimens are horizontal, two-stiped forms, belonging either to *Didymograptus (Expansograptus)* or *Xiphograptus*. The distinction between these genera relies solely on the presence of a virgella on the sicula of *Xiphograptus*, but unfortunately this is commonly not visible in poorly preserved material. The specimens do not exhibit a virgella, but agree well in all other respects with the revised definition of *Xiphograptus svalbardensis* (Archer and Fortey) published by Williams and Stevens (1988). The material from Little Northwest Arm is therefore here referred questionably to that species. *X. svalbardensis* first occurs in mid-Arenig strata of the Cow Head Group (Williams and Stevens, 1988), but there ranges into at least the late Arenig. The association of *Holmograptus* sp. cf. *H. lentus* and *Xiphograptus svalbardensis?* does not permit an unequivocal age determination, but perhaps suggests a broadly similar late Arenig–early Llanvirn age to that deduced for the Strong Island locality.

The graptolite assemblages discussed above are particularly important in light of the biostratigraphic evolution of Middle and Upper Ordovician rocks near the New Bay Fault Zone (Figure 2), a regionally significant structure affecting the Exploits Group, the Lawrence Harbour Formation and the Point Leamington Formation (Helwig, 1969; O'Brien, 1991a). Within this fault zone, Ashgill strata of the *D. complanatus* Zone are separated from uppermost Arenig–lowermost Llanvirn strata of the *D. hirundo* or *D. artus* zones by a multi-unit sequence of folded and faulted beds whose combined structural–stratigraphic thickness is as little as 500 m. Of note is the fact that some distance from the New Bay Fault Zone the pre-*D. complanatus* portion of the Point Leamington Formation alone exceeds this thickness. Nevertheless, within this tract of attenuated stratigraphy, most regionally developed units are present, such as the trace-fossil horizon and lowest polymictic conglomerate lenticle in the lower Point Leamington Formation, the *P. linearis*, *D. clingani*, *D. multidens* (?) and *N. gracilis* intervals from parts of both the upper and lower Lawrence Harbour Formation, the *Holmograptus*-bearing beds of the Strong Island chert and the uppermost volcanic formation of the Exploits Group (see Dec *et al.*, *this volume*). East of the New Bay Fault Zone, recent graptolite discoveries have been made near the base of the uppermost volcanic formation of the Exploits Group, but the fauna has yet to be positively identified and compared with that of the Strong Island chert. Still, it is highly probable that both the Mid and Late Ordovician successions within the fault zone were originally condensed, and that the regional juxtaposition of early Ashgillian and late Arenigian units cannot be solely explained by anomalously large, post-Ashgillian displacements in the New Bay Fault Zone.

St. Josephs Cove Formation, Baie d'Espoir Group

Williams (1991a) previously recorded the presence of Middle Ordovician graptolites from several black shale localities in the Baie d'Espoir Group (Figure 3; Colman-Sadd, 1980). However, he failed to recover any material from Twillick Brook along the river, 50 to 100 m east of the Bay d'Espoir highway (48°4'N, 55°34'W), where black shale occurs near felsic volcanic rocks of the Twillick Brook Member of the Baie d'Espoir Group. He commented that the reason for the absence of graptolites was unclear, as both lithology and lack of intense structural deformation seemed ideal for their preservation. On lithological grounds, he

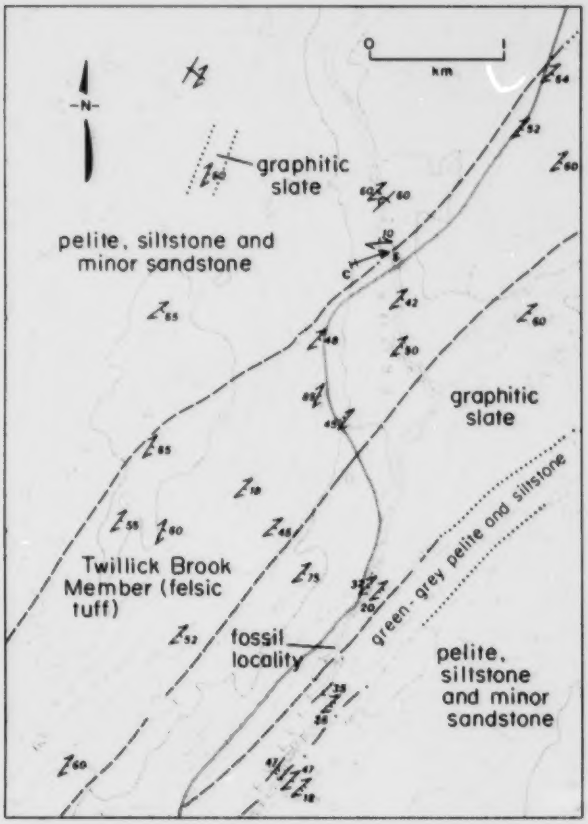
0
1
km
-N-
graphitic slate
pelite, siltstone and minor sandstone
graphitic slate
green-grey pelite and siltstone
Twillick Brook Member (felsic tuff)
fossil locality
pelite, siltstone and minor sandstone

Figure 3. *Location of graptolitic black shale in the St. Josephs Cove Formation of the Baie d'Espoir Group at Twillick Brook, Bay d'Espoir.*

considered that the black shale was identical to the graptolitic unit exposed in neighbouring localities, and similar to the lower part of the Lawrence Harbour Formation, exposed some 100 km distant across the structural width of most of the Exploits Subzone (Figure 1).

During 1991, a few graptolites were found in black shales at the southernmost end of the river section, just below a small waterfall and below a newly built Department of Fisheries cabin. The graptolites are too poorly preserved for identification of any kind, but do appear to be similar to those Caradoc examples found elsewhere in the Baie d'Espoir Group. This would imply a significantly younger age for the black shale than the Llanvirn age determined radiometrically from adjacent volcanic rocks of the Twillick Brook Member (Colman-Sadd *et al., in press*).

CONCLUSIONS

In its type area, the Upper Ordovician Point Leamington Formation contains several graptolite-bearing lenticles of shaly flysch that extend the age range of the Formation from the *D. clingani* Zone of the Caradoc to the *P. pacificus* Subzone (*D. anceps* Zone) of the Ashgill. A paleoenvironmentally unique graptolite assemblage, most probably belonging to the *D. artus* Zone of the Llanvirn, has been recovered from hemipelagic strata presently included in the upper part of the Middle Ordovician (and older ?) Exploits and Wild Bight groups. Black carbonaceous shale of Caradoc age occurs throughout much of northeastern and southern Newfoundland and is present within several major rock groups of the Exploits Subzone.

ACKNOWLEDGMENTS

The authors thank Erwin Wheaton for his assistance in the field and acknowledge, in particular, his keen sense of smell for fossils. This work on graptolites (SHW) and on conodonts (FHCOB) was undertaken under separate paleontological research contracts with the Geological Survey Branch, Newfoundland Department of Mines and Energy. Fieldwork carried out by BHOB was funded by the Canada–Newfoundland Cooperation Agreement on Mineral Development (1990-94).

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GEOLOGY AND GOLD MINERALIZATION OF THE DUDER LAKE GOLD SHOWINGS, EASTERN NOTRE DAME BAY, NEWFOUNDLAND

R.A. Churchill and D.T.W. Evans[1]
Department of Earth Sciences, Memorial University of Newfoundland
St. John's, Newfoundland A1B 4J6

ABSTRACT

Structurally controlled, gabbro-hosted gold mineralization occurs at Duder Lake, northeastern Newfoundland. Discovered in 1988 by Noranda Exploration Company Limited, it is one of several recent gold finds in rocks of the eastern Dunnage Zone.

The mineralization is confined to wall-rock alteration adjacent to brittle low-angle Riedel shears. The gold occurs as fine intergrowths with pyrite, arsenopyrite, and minor sulphosalts, which formed contemporaneously with titaniferous, siliceous, and carbonate alteration. Mineralization style is manifested as disseminated, semi-massive, and stringer sulphides.

The Corvette and Goldstash prospects, as well as the Flirt showing, delineate a 4-km-long shear zone that hosts significant concentrations of gold. A third prospect, Stinger, was inferred to be related to a splay originating from the main structure.

The Duder Lake gold showings are hosted by late gabbroic dykes and sills that intrude rocks of the Davidsville and Botwood groups. This mineralization style is similar to other gabbro-hosted gold showings in the eastern Dunnage Zone.

INTRODUCTION

A study of the gold occurrences hosted by Siluro-Ordovician sedimentary and gabbroic rocks at Duder Lake was initiated in the summer of 1991, in conjunction with a regional gold-metallogeny project carried out by the Newfoundland Department of Mines and Energy (see Evans, *this volume*). This study, which forms the basis of a M.Sc. thesis by the first author, will focus on structural controls of mineralization and alteration, details of the alteration associated with gold mineralization, possible mineralizing fluid sources based on geochemical and isotopic studies, and affinity of the gabbroic bodies closely associated with mineralization at Duder Lake. This work will include grid and trench mapping, diamond-drill-core logging, and rock sampling for geochemical and isotopic analyses.

Duder Lake (Figure 1) is located approximately 3 km east of Birchy Bay, Notre Dame Bay. The area is readily accessible by logging roads from Birchy Bay that lead west and south of Duder Lake, and from Loon Bay that lead to the eastern and northern parts of the study area.

The gold showings at Duder Lake are located on a 9-km-long north-trending peninsula, which separates Duder Lake from Rocky Pond to the east and from Ten Mile Lake to the south. The peninsula is characterized by low relief, and elongate ridges that appear to be controlled by the geological structures that transect the peninsula. The southern part of the peninsula is heavily forested and contains numerous small bogs and dense alder beds; the northern areas have been cut over by recent logging.

The morphology south of Duder Lake, Rocky Pond, and Ten Mile Lake reflects regional northeast structural controls and bedrock grain. The region is covered by a thin till veneer and locally thick glaciofluvial deposits.

REVIEW AND REGIONAL SETTING

Duder Lake is located in the eastern Dunnage Zone of the Newfoundland Appalachians (Williams, 1979). The Dunnage Zone records the evolution, and subsequent destruction, of the Paleozoic Iapetus Ocean. The geological history of the Dunnage Zone is subdivided into two stages: pre-accretion and post-accretion.

Pre-accretion events include the development of island arcs and back-arc basins during the Cambrian to Middle Ordovician. These systems recorded both pre- and syn-accretion sedimentation, which is manifested as distal turbiditic sedimentation. Cessation of volcanism during the

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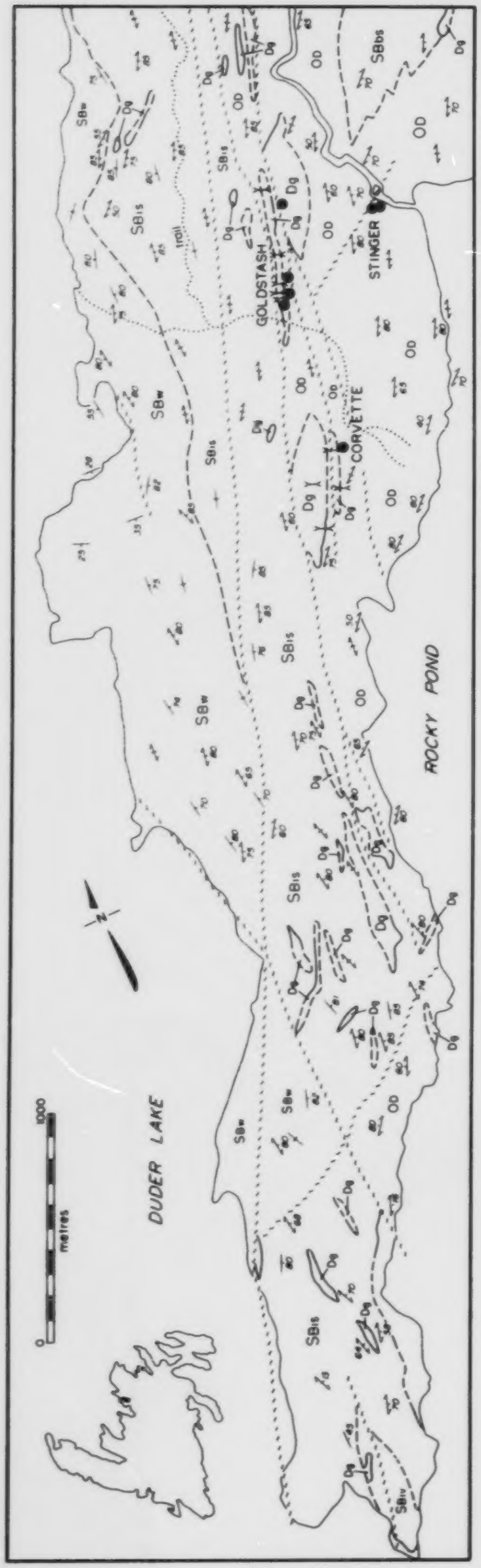
[1] Mineral Deposits Section, Geological Survey Branch, Newfoundland Department of Mines and Energy



LEGEND

DEVONIAN

Dg — *Medium- to coarse-grained gabbro, diorite, and diabase, locally leucoxene rich. Primarily igneous layering sometimes present as well as pegmatitic textures in larger bodies. Usually dark-green but when altered, exhibits a distinct reddish-brown oxidation surface*

SILURIAN

Botwood Group

Wigwam Formation

SBw — *Red and green sandstone, and minor siltstone. Massive to locally laminated with crossbedding, and graded bedding. Siliceous and micaceous moderate cleavage developed locally*

Lower Unit

SBlv — *Pillowed basaltic flows and breccias; vesicular*

SBls — *Undivided olive-green, green and red, and red siltstone, minor sandstone, and slate. Thinly interbedded with local massive units up to 25 m in thickness. Moderate to strong penetrative cleavage developed*

ORDOVICIAN

Davidsville Group

OD — *Dark-grey, green, and black slate with minor siltstone and fine-grained sandstone. Thinly interbedded and exhibiting a strong penetrative slaty cleavage*

Contact (defined, approximate) ——— – – –

Faults ﹏﹏

Strike and Dip (dipping, vertical) ⊥ +

S_1 Schistosity (dipping, vertical) ⇀ ⇌

Drill Road ....

Trench ———

Diamond Drill Hole ●

Figure 1. *Geology and location of gold prospects in the Duder Lake area, showing location of drillholes and trenches.*

Middle Ordovician corresponded with the emplacement of the Taconic allochthons. Continued closure of Iapetus resulted in flyschoid-sediment deposition in fault-bound basins of the east-central Dunnage Zone.

The Davidsville Group (cf. Kennedy and McGonigal, 1972) is a thick sequence of distal back-arc turbidites, which were fed from the older island-arc systems to the west, and were deposited on allochthonous oceanic basement delineated by rocks of the Gander River Ultrabasic Belt (Blackwood, 1982). This distinct belt of northeast-trending ultramafic rocks have been formally termed the Gander River Complex by O'Neill (1991).

Shelf-facies rocks of the Gander Zone pass conformably westward into rocks of the Dunnage Zone in north-central Newfoundland. The Dunnage Zone is interpreted to be regionally allochthonous upon rocks of the Gander Zone (Colman-Sadd and Swinden, 1984).

Post-accretionary events include deposition of the Silurian Botwood and Indian Island groups. During the Late Ordovician to Middle Silurian, regional-scale transcurrent discontinuities were either activated or reactivated causing the development of pull-apart basins (Szybinski *et al.*, 1990) and crustal anatexis resulting in widespread epicontinental-style volcanism (Coyle and Strong, 1987). Basin infilling was characterized by the deposition of fluviatile- to shallow-marine facies accompanied by subaerial volcanism forming the Botwood and Indian Island groups.

The Botwood Group comprises grey to red micaceous sandstone and siltstone, minor fossiliferous calcareous beds, and conglomerate (Williams, 1964). The Indian Islands Group consists of phyllitic slates, quartzitic and calcareous sandstone, thin limestone lenses, and conglomerate (Baird, 1958). Patrick (1956) also identified minor felsic volcanic rocks intercalated with this unit.

The intense deformation and plutonism prevalent in the eastern Dunnage Zone, originally thought to be related to Acadian deformation, is now considered to have resulted from a Silurian orogenic event (Dunning *et al.*, 1990).

Rocks of both the Botwood and Davidsville groups record the deformation as a regional, northeast-trending penetrative cleavage (S_1), which is axial planar to isoclinal folds. Cleavage in the Davidsville Group is slaty and relatively well developed; in contrast, the Botwood Group does not exhibit this slaty characteristic and cleavages are only well developed, in local instances, where cleavage is strongly developed by regional shearing. Second-phase deformation (S_2) is typified as small conjugate kink bands and minor open to moderately tight southward-plunging folds that fold the F_1 axial-planar cleavage (Blackwood, 1982). Metamorphism is confined to lower-greenschist facies.

Intruding both the Davidsville and Botwood groups are small, fine- to coarse-grained gabbroic sills and dykes, which appear to be most prevalent at, or near, the contact between the two groups. These gabbroic bodies are generally undeformed except where they have been sheared by movement along north- to northeast-trending brittle shear zones. These gabbroic rocks are observed elsewhere in the eastern Dunnage Zone. Mafic plutonic rocks were observed to intrude both Davidsville and Botwood groups in the Gander River area and form part of the Mount Peyton Intrusive Suite (Blackwood, 1982).

PREVIOUS WORK

The first recorded geological survey of north-central Newfoundland was done by Murray and Howley in 1881. This was a regional mapping investigation, in which the geology was examined in eastern Notre Dame Bay, the Gander River, and Gander Lake areas. Several regional studies have encompassed the study area.

A geological survey of the Notre Dame Bay area was undertaken by Kranck (1952). Part of the study focussed on the geology in the vicinity of Loon Bay. Kranck (1952) noted six different units comprising the stratigraphy of this area as well as the strong deformation present. Patrick (1956) examined the geology in the Comfort Cove map area (NTS 2E/7) and assigned rocks in the Duder Lake area to the Springdale Group. He defined two sedimentary units in the vicinity of Duder Lake. The westernmost unit was described as red, brown, grey, and green quartzite and quartzitic sandstones. The unit to the east was described as red, brown, grey, and green slates, argillites and shales. In addition, small gabbroic dykes and sills of Devonian age were recorded outcropping along the southeast shore of Duder Lake. Williams (1964) undertook a study to compile the geology, complete initial geological mapping, and to investigate the mineral deposits in northeast Newfoundland. In the Duder Lake area, he identified two different units: a Silurian-aged unit composed of red and grey micaceous sandstone and siltstone, conglomerate, shale, coralline shale, and limestone; and an Ordovician-aged unit consisting of grey to black slate and siltstone, greywacke and minor volcanic rocks. The Silurian-aged unit was mapped as part of the Botwood Group whereas the Ordovician-aged unit was determined to be part of the Gander Lake Group.

Several regional stratigraphic, tectonic, and structural studies have focussed on the Silurian and Ordovician stratigraphy of north-central Newfoundland. Most notable is the work of Twenhofel and Shrock (1937); Heyl (1937); Williams (1967); Kay (1967); Horne and Helwig (1967); Dean (1977, 1978); Nelson and Casey (1978); Arnott *et al.* (1985) and Goodwin and O'Neill (1991).

Since 1980, the eastern Dunnage Zone has become the focus of extensive gold exploration after the discovery of quartz-vein-hosted gold mineralization in the Jonathan's Pond area (Blackwood, 1982).

The Newfoundland Department of Mines and Energy, in 1988, released the results of a regional lake-sediment geochemical survey, which yielded anomalous Au, Sb, and As over a wide area in the eastern Dunnage Zone (NTS 2E, Davenport and Nolan, 1988). The highest concentrations of As and Sb were located at Duder Lake, Rocky Pond, and Ten Mile Lake. This prompted exploration companies, such as Noranda Exploration Company Limited, to stake ground in the immediate area since the Au–As–Sb association usually accompanies Au mineralization.

As a result of prospecting and soil geochemical surveys, several significant gold discoveries were made in rocks of the Davidsville and Botwood groups (Green, 1989; Tallman, 1990).

LOCAL GEOLOGY

The Duder Lake area encompasses the contact between the Botwood and Davidsville groups. Mapping by Noranda geologists in 1988 and 1989, and by the authors in 1991, have identified outcrops of both the Botwood and Davidsville groups. These units are present as north-trending belts, which divide the area separating Duder Lake from Rocky Pond (Figure 1). The Botwood Group is subdivided into two units: the Wigwam Formation (Dean, 1977), and a sequence of undivided olive-green, greenish-red sandstones and siltstones informally named the 'lower unit'.

Intruding these sedimentary units are gabbroic dykes and sills, which are important to mineralization.

Deformation within the Duder Lake area is recognized as a well-developed, penetrative cleavage (S_1), which parallels the regional fabric (Plate 1) and is axial planar to F_1 folds produced during the regional deformation. Lower hemisphere projections of the structural data (Figure 2) indicate that cleavage orientation is northeast trending and has variable inclinations from shallow dipping to vertical (Figure 2a). The bedding plane data (S_0) is observed to be quite varied in orientation and dip. However, a π-girdle can be constructed through the data points, which infers that the fold-axis trends in a northerly direction (Figure 2b).

Comparing the two plots, it is apparent that the cleavage is axial planar to the folding, which reaffirms the regional structural relationships.

BOTWOOD GROUP

Wigwam Formation

This unit is exposed along the western and eastern shorelines of Duder Lake. It is characterized by massive and locally laminated, maroon micaceous sandstone and siltstone containing minor conglomeratic lenses and beds. These rocks contain abundant shallow-water sedimentary structures such as convoluted bedding, dessication cracks, crossbedding and ripple marks. Bedding planes dipping steeply toward the east and south and the variations in attitudes can be attributed to broad open-folding.

Plate 1. *Penetrative axial-planar cleavage representative of the* S_1 *deformational fabric at Duder Lake.*

Deformational fabrics are not well developed, but intense S_1 cleavage is developed and locally parallels the regional fabric.

Lower Unit

This informally termed sedimentary unit forms a second north-trending belt to the east of the Wigwam Formation, separating the Wigwam Formation and the Davidsville Group. The contact with the Wigwam Formation appears to be fault-modified but conformable. Although classified as part of the Botwood Group, the sedimentary rocks of this unit also have features similar to those of the Davidsville Group.

This unit is divided into two subunits—one sedimentary and the other dominated by mafic vesicular pillow basalts, flows, and breccias. Williams (1964) observed that the Botwood Group contained red, green, and purple amygdaloidal lavas, agglomerates, siliceous tuff, and minor sandstone, which may be intercalated with terrestrial micaceous sandstones and siltstones of typical Botwood Group affinity.

The sedimentary unit comprises locally micaceous, olive green, greenish-red, and red siltstone, minor sandstone, and slate. These rocks are thinly interbedded but contain local massive units of undivided sandstone and siltstone up to 25 m in thickness. Detrital mica content is observed to decrease eastward toward the Davidsville Group. This unit exhibits a moderate to strongly developed cleavage, which becomes slatey toward the contact with the Davidsville Group. The contact between this unit and the Davidsville Group is inferred to be a fault-modified gradational contact.

To summarize, it would appear that the lower unit records a transition between the Wigwam Formation and the Davidsville Group, which has been observed elsewhere in the eastern Dunnage Zone (Blackwood, 1982). The variable detrital mica contents observed in rocks of the Davidsville and Botwood groups reflect the different provenance of both

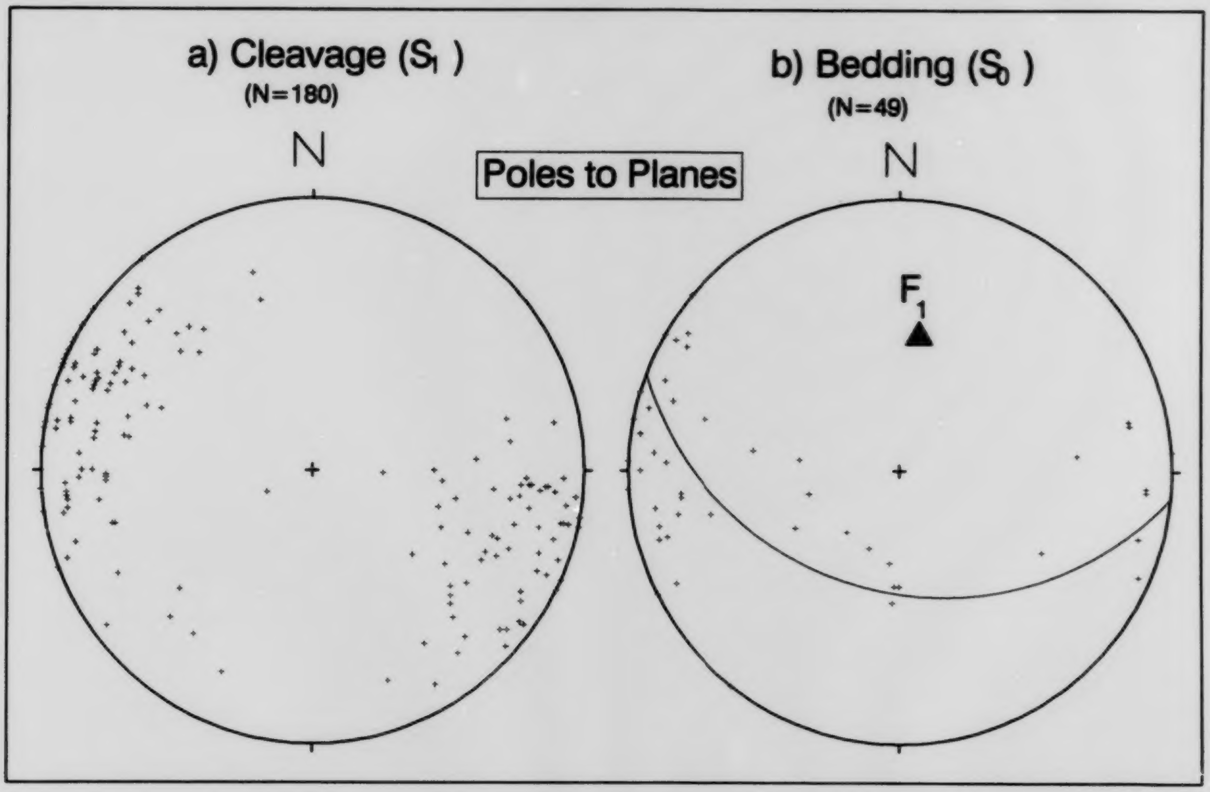


Figure 2. *Lower hemisphere projections for structural data collected from Duder Lake. (a) cleavage, (b) bedding. Data is presented as poles to planes (refer to text for discussion).*

groups. Sedimentary rocks of the Davidsville Group were derived from the island-arc systems to north and west. Evolution of the basin infilling denoted a shift in provenance during earliest Silurian times. An easterly mica-rich source region supplied the material that formed the Botwood Group. Blackwood (1982) suggests that this source was the high-grade metamorphic terrane of the Gander Zone lying to the east.

DAVIDSVILLE GROUP

The Davidsville Group in the Duder Lake area comprises a monotonous sequence of dark-grey, green, and black slate, associated with minor siltstone, fine-grained sandstone, and minor carbonaceous sandstone and siltstone. These rocks are thinly interbedded and exhibit strong penetrative slaty cleavage, which is axial planar to the F_1 folds. This fabric is locally crenulated.

GABBROIC DYKES AND SILLS

Numerous small gabbroic bodies intrude all units in the Duder Lake area. The bodies are typified by dykes and sills—some of which still possess primary igneous layering.

The gabbroic rocks have varied mineralogical compositions and two distinct gabbro types can be delineated: melanocratic and leucocratic. Melanocratic gabbros appear to be more refractory and have an abundance of Fe- and Ti-rich minerals. The leucocratic gabbros contain less of these minerals, thus yielding a lower colour index (Plate 2).

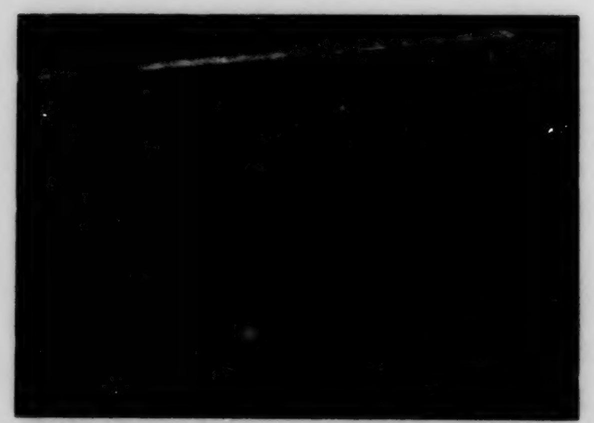


Plate 2. *Compositional and textural variety observed in the gabbro from drill core.*

Preliminary results indicate that only the melanocratic gabbros that intrude the Davidsville Group, host significant mineralization.

The textures of the gabbroic intrusions are variable ranging from extremely coarse grained in the sills to fine grained (almost aphanitic) in the dykes. Coarse pegmatitic pods of gabbroic material occur in some of the coarser grained gabbros. The gabbros are typically undeformed except when found proximal to the north-trending shears. In such cases, the gabbros exhibit well-developed schistosity. The gabbros appear to be related to the lineaments transecting the area and are shown in Figure 1. A two-stage model is derived for their emplacement and subsequent deformation: 1) Activation of the north-trending structures may have supplied enough frictional heating at depth to allow for small-scale crustal melting giving rise to the gabbroic rocks; and 2) subsequent movement of the structures, either during activation or later reactivation, caused the deformation observed in these rocks.

GOLD MINERALIZATION

In the Duder Lake area, exploration programs were carried out by Noranda Exploration Company Limited and their joint-venture partner Noront Resources, during 1988 and 1989. As a result of prospecting and regional till- and soil-geochemical surveys, three gold prospects (Figure 2) and one gold showing were outlined in the Duder Lake area—Goldstash, Corvette, Stinger, and Flirt (not shown on figure), respectively. All four mineralized zones are spatially related to a south-trending, 4-km-long shear zone. Assays from the Goldstash Showing drill-core yielded 4.2 g/t over 7.8 m, of which 2.8 m of this interval ran 6.4 g/t. Trench sampling from the same showing yielded an average grade of 12.8 g/t over a 2.6 m interval.

Evans (1991, *this volume*) defined several styles of gold mineralization after documenting gold showings in the eastern Dunnage Zone. This classification nomenclature will be utilized in the discussion on the gold mineralization at Duder Lake.

Goldstash, Corvette, and Flirt (Type 1)

The Goldstash and Corvette showings at Duder Lake are both hosted by gabbros that intrude rocks of the Davidsville Group. The style of mineralization at these showings is structurally controlled and is termed Type-1 Disseminated Mesothermal Mineralization (gabbro-hosted).

Gold mineralization is largely confined to the wall rock and is controlled by brittle structures, which appear to be low-angle Riedel shears that trend 000 to 010° (Green, 1989) and originate from the regional shears. The highest gold grades are found in wall-rock mineralization.

Alteration assemblages adjacent to the shears delineate an alteration halo, which decreases in intensity away from the shear. This halo is comprised of three zones:

1) Alteration within the shear is intense Fe-oxidation with accompanying Fe-carbonate–carbonate–quartz-alteration phases. In addition, small shear-parallel, barren, quartz–carbonate veins are present, locally. Extensive weathering of this lateration style combined with the jointing pattern, produces a 'blocky' alteration pattern similar to that described by Evans (1991) from Clutha (Plate 3). Mineralization is confined to the sheared gabbroic rock and comprises finely disseminated pyrite and arsenopyrite.

Plate 3. *Recessive weathering and accompanying 'blocky' alteration pattern of mineralized shear structure as observed in trench wall.*

2) The next alteration assemblage found immediately adjacent to the shear is delineated by intense wall-rock alteration, where the gabbroic host is almost completely replaced by silica and carbonate. Such alteration attests to the volume of silica- and carbonate-rich fluid that must have been localized and transported by the shear. This is the zone that contains the most significant economic concentrations of gold, pyrite, and arsenopyrite mineralization.

3) The third alteration assemblage is characterized by quartz–sericite–leucoxene. This zone passes gradationally into relatively fresh gabbro. Mineralization in this zone occurs as very fine disseminations of pyrite and arsenopyrite in low concentrations.

Leucoxene is a common mineral phase present in all the showings. At present, its relationship with unaltered gabbro undergoing hydrothermal alteration is undetermined. However, intense leucoxene alteration (almost 90 percent) can be observed in drill core and in the trench at the Corvette Showing. Such abundance of a TiO_2-rich amorphous mineral phase would imply that the gabbro was initially rich in ilmenite or rutile.

The size of the alteration haloes are proportional to the size of the shear that produces them. The zonation depicted in Figure 3 was the result of the development of a 0.75-m-

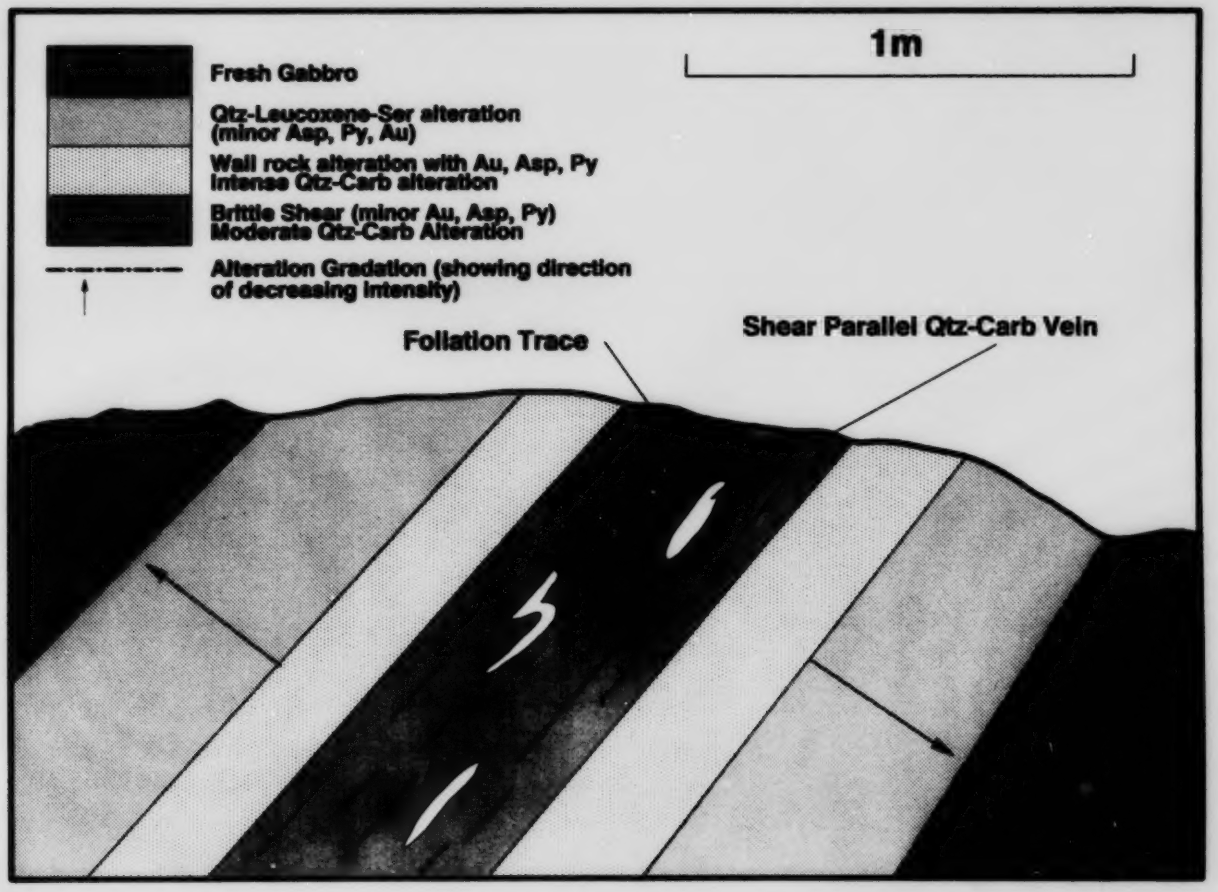

1m
Fresh Gabbro
Qtz-Leucoxene-Ser alteration (minor Asp, Py, Au)
Wall rock alteration with Au, Asp, Py Intense Qtz-Carb alteration
Brittle Shear (minor Au, Asp, Py) Moderate Qtz-Carb Alteration
Alteration Gradation (showing direction of decreasing intensity)
Foliation Trace
Shear Parallel Qtz-Carb Vein

Figure 3. *Schematic diagram showing the features observed in trench from the Goldstash Showing. Note the alteration halo and accompanying mineralization surrounding the brittle shear.*

wide shear. In this instance, the entire zonation spanned a distance of approximately 1 to 1.5 m on either side of the shear.

Stinger Prospect (Type 2)

Evans (1991) stated that the Stinger Prospect comprises sediment-hosted auriferous quartz–carbonate veins with accompanying carbonatization and silicification developed in carbonaceous fine-grained slates and siltstones. Further work is required on this showing since it was not accessible during the time spent at Duder Lake (high water levels did not allow the first author to investigate the mineralized zones that outcrop in Rocky Pond River).

GEOCHEMISTRY

Preliminary data, shown in Figure 4, illustrate some of the bulk chemical characteristics of the fresh, altered, and mineralized gabbros. The distinction between these three gabbro types are based on their field appearance. Fresh gabbros contain no appreciable alteration or mineralization. Altered gabbros are representative of the alteration zonations observed at sites of mineralization and possess little or no pyrite–arsenopyrite mineralization. The mineralized gabbros are typically intensely silicified and carbonatized and contain appreciable concentrations of sulphides. The plot of Log Cr versus Log Ni (Figure 4a) does not discriminate between any of the three gabbro types, implying no appreciable Cr or Ni was exchanged with the mineralizing fluids. As a result, the Cr and Ni concentrations probably reflect pre-alteration concentrations.

Ternary plots of Cu–As–Zn and Cu–Pb–Zn (Figure 4b) characterize bulk chemical changes that occurred during alteration and mineralization. The mineralized gabbros show a bimodal distribution of arsenic from values, which correspond to original fresh gabbro to concentrations in excess of 40,000 ppm. This would suggest that As was being brought into the system from an exterior source, and its concentration does not reflect original compositions.

Further geochemical work will better characterize the geochemical changes observed in the system and should lead to some inference of fluid composition and/or source.

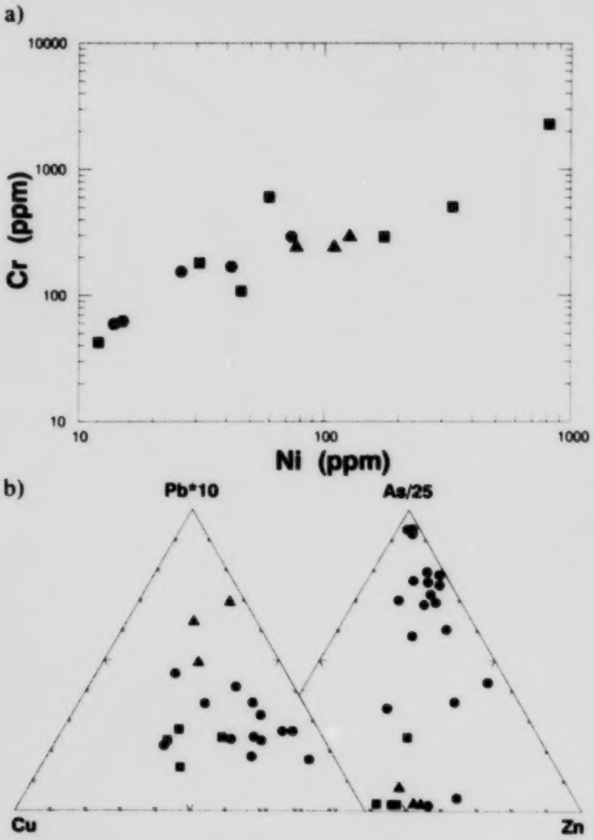


Figure 4. *Geochemical plots of fresh gabbros (squares), altered gabbros (triangles), and mineralized gabbros (circles). a) Plot of Log Cr versus Log Ni. b) Ternary plots of Cu–Pb–Zn and Cu–As–Zn.*

DISCUSSION

The styles of mineralization observed at Duder Lake are comparable to similar gold mineralization found elsewhere in the eastern Dunnage Zone. The preliminary investigation at Duder Lake implies that alteration and mineralization is strongly controlled by structure. Evans (*this volume*) suggests that the style of mineralization observed at Duder Lake is reminiscent of mesothermal-style mineralization. Although valid for the structural controls on mineralization, lode-gold mineralization has not been observed in the vein networks at the Goldstash Showing. It appears that mineralization is best developed in wall-rock alteration, which is accompanied by intense silicification and carbonatization.

Since mineralization is confined to gabbros, which have been sheared by second- and third-order structures, it would imply that they must be rheologically suitable as a site for mineralization. Compared to the surrounding rocks, the gabbros would be enriched with respect to Mg and Fe. This mechanism is not suitable to explain mineralization at the Stinger Prospect, which is sediment-hosted. The carbonaceous nature of the sandstones and siltstones at the Stinger Prospect may control fluid precipitation.

Barley and Groves (1990) state that a degree of disequilibrium is produced when auriferous fluids rich in silica and carbonate interact with rocks having a high Fe/(Fe+Mg) ratio. As a result of disequilibrium, the system becomes unstable and precipitates pyrite, arsenopyrite, and gold. A similar situation can be ascribed for the Stinger Showing, as the organic carbon found in the rocks would lead to chemical reduction if the fluid was originally oxidized (Stokes *et al.*, 1990).

The silicification and carbonatization observed accompanying deformation can be likened to the brittle-ductile transition zone of mesothermal deposits, which possess a vertical gold zonation (Colvine *et al.*, 1988).

Presently, drill-core and trench observations infer that, if the model is valid, only the uppermost part of the vein systems are being seen (i.e., the uppermost part of the brittle-ductile transitionary zone).

The sedimentary packages of the eastern Dunnage Zone have been affected by the Silurian orogenic event resulting in high-heat flow, intense brittle and even ductile transpressive structure development, which may reach deep-crustal levels, and an abundant supply of H_2O- and CO_2-rich fluids, which would be focussed up along these major crustal discontinuities ultimately culminating in gold mineralization at upper crustal levels (Kerrich, 1989; Sibson *et al.*, 1988).

Future investigations in the Duder Lake area will concentrate on the structural constraints placed on mineralization and alteration, the affinity of the gabbroic bodies, better constraint on the Davidsville–Botwood contact, and geochemical and isotopic evidence to deduce a genetic model for these showings. The completion of these tasks will hopefully give insight into the regional significance of the Duder Lake showings and analogous gold and base-metal showings elsewhere in the eastern Dunnage Zone in an effort to develop a regional exploration model for this region.

ACKNOWLEDGMENTS

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GEOLOGY OF THE PORCUPINE STRAND BEACH PLACERS, EASTERN LABRADOR

M. Emory-Moore and J. Meyer[1]
Centre for Cold Ocean Resources Engineering (C-CORE)
Memorial University of Newfoundland, St. John's, A1B 3X7

ABSTRACT

The Porcupine Strand, in eastern Labrador, consists of over 40 km of sandy beach, known to contain accumulations of heavy mineral sands (iron–titanium, garnet). The 1991 field program included a small, drill program carried out by Cominco Limited, channel sampling and geophysics, with emphasis placed on the North Strand. The best economic potential lies in the modern and raised beach sands, which developed during a time of relatively slow coastal emergence and low sediment supply. The reworking of a finite supply of sediment has produced a relatively high-grade placer deposit of limited extent.

INTRODUCTION

The placer potential of modern and raised beach sands of the Porcupine Strand, eastern Labrador, is presently being investigated by Cominco Limited, the Newfoundland Department of Mines and Energy and Centre For Cold Ocean Resources Engineering (C-CORE). The potential economic value of these placer occurrences lies principally in the iron and titanium (Fe–Ti) content of the magnetite, hematite and ilmenite. The purpose of this paper is to provide a review of the results from the 1991 field season and to advance a depositional model for placer development in the area.

The Porcupine Strand consists of over 40 km of sandy beach tracts separated by low-relief rock headlands (Figure 1). The most prominent headland is Cape Porcupine, which separates the North Strand from the South Strand. Beach slopes are generally low (< 3°) with the intertidal zone ranging from 50 to 100 m in width and the backshore, from 10 to 40 m in width (Plate 1). All beaches are cut into a broad coastal plain, which abuts a more rugged mountainous terrain to the west of the Strand. The coastal plain is deeply incised by the North River, the single major river in the area.

The placer potential of the Porcupine Strand beaches was first investigated by Emory-Moore and Meyer (1991). Based on 10 test pits, 50 to 100 cm in depth, an estimated 33,600 tonnes of Fe–Ti concentrate and 9,150 tonnes of garnet concentrate were delineated over 1 km of the beach; the overall thickness and lateral extent of the mineralized sand was not established.

Based on field observations, the South Strand beaches (both modern and raised) were noted to have far lower heavy mineral concentrations, and presumably less mineral potential. Thus, the following discussion is limited to placer occurrences on the North Strand.

STUDY AREA

Bedrock Geology

The study area falls within the northeastern portion of the Precambrian Grenville Province and is underlain by paragneiss, granitic orthogneiss, and granulite-facies gabbroic, monzonitic and granitic rocks (Gower, 1984). Immediately to the west and south of the Porcupine Strand there are numerous thick, folded gabbronorite–monzonite layered intrusions (Gower *et al.*, 1981). These have not been prospected but given their layered nature they almost certainly contain fractionated Fe–Ti oxides.

Quaternary Geology

During the Quaternary, glacial ice extended onto the continental shelf of Labrador, into present water depths of approximately 600 m (Josenhans *et al.*, 1986). The extent of late Wisconsinan ice is not well defined in the study area; some workers believe ice extended into the present offshore (Josenhans *et al.*, 1986) whereas others believe that the ice margin was close to the present coastal limits (Rogerson, 1977). It is generally agreed that ice flow was toward the coast,

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[1] Mineral Deposits Section, Geological Survey Branch, Newfoundland Department of Mines and Energy

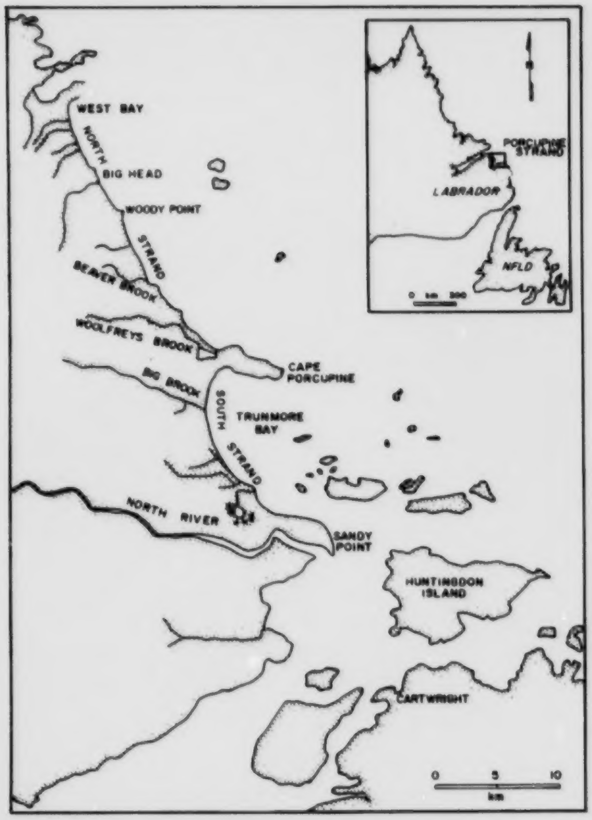

Figure 1. *Location map and place names, Porcupine Strand, eastern Labrador.*

Plate 1. *Aerial view of the North Strand beach (tract).*

spreading east from a divide that occupied the central part of the Labrador peninsula and that deglaciation commenced almost 12,000 years BP (Vincent, 1989).

Thick Quaternary sequences underlie the coastal lowlands of the study area (Figure 2). An extensive suite of valley and plain sandurs cover much of the lowlands and overlie a thick sequence of glaciomarine mud (Rogerson, 1977). Postglacial isostatic rebound in the study area resulted in a drop in relative sea level of over 60 m (Figure 3; Rogerson, 1977). Sea level dropped at a rate of approximately 1.2 to 2.4 m/100 years from 10,000 to 6,000 years BP. A minor transgression in sea level occurred after 5,600 years BP and has since been followed by a slow (0.18 m/100 years) continuous drop in sea level (Rogerson, 1977).

Oceanographic Regime

During the ice-free season, from June to November, the North Strand is fully exposed to the Labrador Sea where wave heights can exceed 9 m and over fifty percent of the time do exceed 1.5 m (Walker, 1984). Tides in the study area are semi-diurnal with tidal range of 1.2 to 3.0 m. Ebb and flood tidal velocity is 10 cm/sec (Canadian Hydrographic Service, 1988).

METHODS OF SAMPLING AND ANALYSIS

Forty-five holes were drilled in the modern beach complexes of the North Strand using a vibra-core drill. In all cases, drill penetration was limited to less than 2 m due to the presence of gravel and clay horizons and in some areas, frozen ground. A composite sample weighing from 1 to 5 kg was recovered from each drillhole.

Forty-seven channel samples were collected from trenches in the modern beach complexes of the Strand and, where possible, exposed sections of raised beach terraces. Near Beaver Brook (Figure 1) several beach profiles were measured and detailed sampling of each undertaken.

Twelve graduated sediment rods were driven into the beach at selected sites and will be used to monitor net sediment erosion and/or deposition. A loosely fitting washer was slipped over the rod and placed on the surface of the beach. At the end of the observation period, the distance from the washer to the original sand surface, as measured on the graduated rod, will provide a measure of net erosion. The distance from the washer to the final sand surface will indicate the amount of subsequent backfilling. Magnetic surveys of the modern beach and raised beach terraces were undertaken using an Omni magnetometer. A total of 49 transects were completed, 10 of which included repeated surveys using different sensor heights.

Geochemical analyses of the samples are being carried out by the Newfoundland Department of Mines and Energy using inductively coupled plasma spectroscopy. Detailed textural and mineralogical analysis of select samples will be undertaken as part of two undergraduate Honours theses at Memorial University.

COASTAL SEDIMENT SUITE

The coastal sediment suite of the North Strand is composed of three sediment types: glaciomarine muds, outwash sands and gravels, and mineralized beach sands. A brief description of each is provided below.

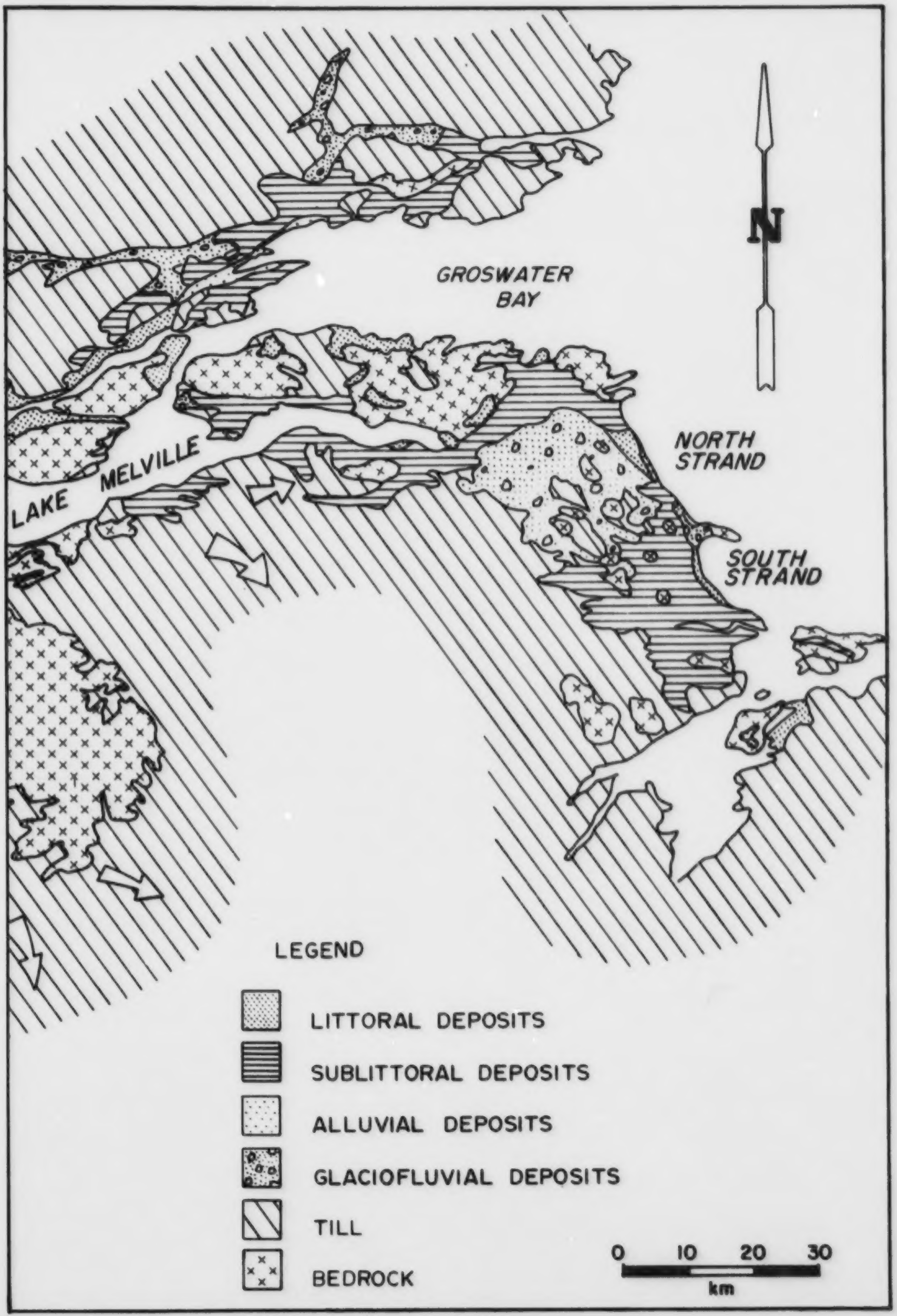
GROSWATER BAY
NORTH STRAND
SOUTH STRAND
LAKE MELVILLE
N
LEGEND
LITTORAL DEPOSITS
SUBLITTORAL DEPOSITS
ALLUVIAL DEPOSITS
GLACIOFLUVIAL DEPOSITS
TILL
BEDROCK
0
10
20
30
km

Figure 2. *Surficial geology (from Fulton, 1986).*

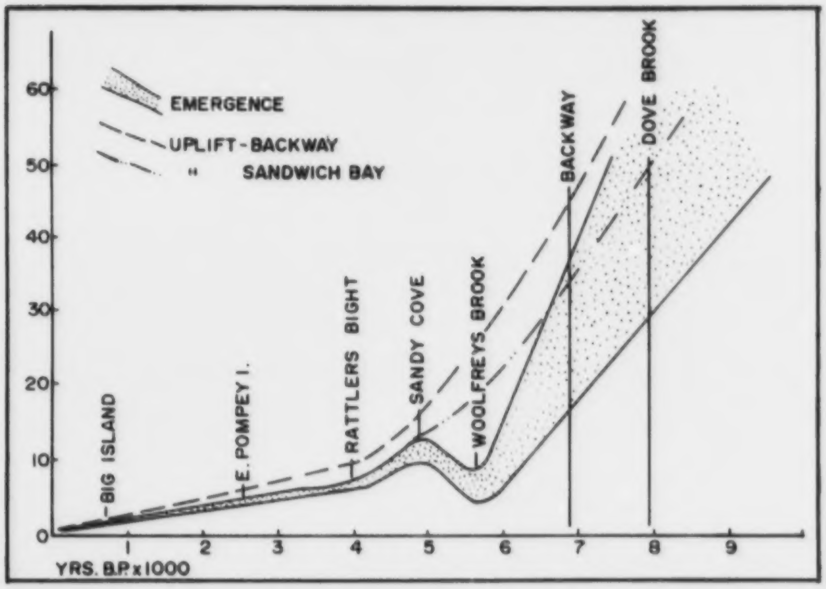

Figure 3. *Postglacial sea-level history (from Rogerson, 1977).*

Glaciomarine Mud

Thick sequences of glaciomarine mud underlie the coastal lowlands of the North Strand, and are well exposed in the modern coastal bluffs, (up to 30 m high). The modern beach and most low-elevation terraces are cut into the relatively resistant bluffs, (with the exception of Woody Point; Figure 4). Glaciomarine muds underlie the modern beach complex at depths of 1 to 2 m. The muds are composed of a silty clay having isolated dropstones, and are generally massive and prone to slumping. The glaciomarine muds were not examined in any detail as they hold no placer potential.

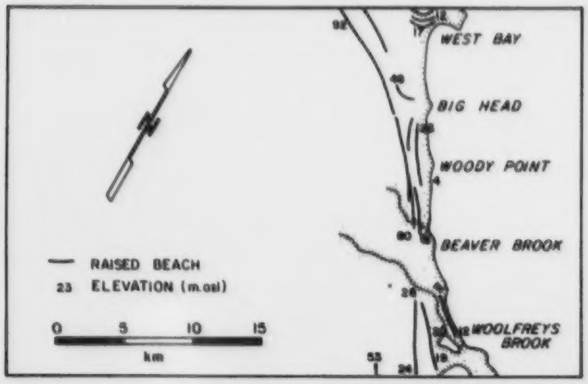

Figure 4. *Distribution of raised terraces (modified from Rogerson, 1977).*

Outwash Sands and Gravels

Outwash sands and gravels overlie the glaciomarine muds and form much of the surficial cover of the North Strand coastal lowlands. Terraces cut into outwash sands and gravels occur at various elevations, including 24 to 28, 48, 61, and 80 to 92 m; in some cases, the terraces can be traced for distances greater than 20 km (Rogerson, 1977; Figure 4). The sand and gravel unit is generally quite thin in coastal exposures (<3 m) except around Woody Point and Woolfreys Brook where it exhibits thicknesses of greater than 5 m.

The outwash sediments exposed in the coastal bluffs are generally composed of moderately sorted medium to coarse sands having interbedded gravels (Plate 2). Sedimentary structures include large and small trough and planar tabular crossbeds, isolated scour fills and low-angle inclined stratification. Outwash sediment is characterized by thin (< 3 mm) heavy mineral laminations, and in many places the sediment is cemented and heavily iron stained. Deflation lags occur on non-vegetated surface exposures.

At higher elevations the outwash sediment becomes coarser grained and less sorted (Rogerson, 1977). Some littoral reworking and redistribution of outwash sediment has occurred on the terrace surfaces.

Mineralized Beach Sands

Mineralized beach sands are found both in low-elevation terraces (<10 m) and on the modern beach complex. The most extensive exposure of mineralized terraces occurs at Woody Point where three terraces were identified (Figure 5). The uppermost terrace (4 m above high tide) abuts outwash sands and is approximately 1 km long and between 25 and 30 m wide. The terrace, at 1.8 m above the high tide level, extends for approximately 5 km and is about 5 to 10 m wide. In many areas, only traces of the lowermost terrace (<0.5 m above high tide) have been preserved.

Two, well-developed mineralized terraces occur immediately north of Beaver Brook (Figure 5). The upper terrace is cut into glaciomarine muds and is approximately 1 km long and 28 m wide. The lower terrace is partially

Plate 2. *Mineralized terrace sands.*

eroded. Low-elevation mineralized terraces (1 to 2 m above high tide level) also occur in some small brook outlets (Figure 5).

The terrace sands are generally composed of subhorizontal planar laminations–beds of light-coloured quartzofeldspathic minerals, alternating with dark-grey, well-sorted heavy mineral layers (Plate 2). The heavy mineral layers are typically 0.3 to 2.0 cm thick, although a 1-m-thick section of massive heavy mineral sands was encountered in a mineralized terrace, approximately 100 m north of Beaver Brook. Heavy mineral laminations/beds constitute from 30 to 80 percent of each terrace and have a slight decrease in occurrence with an increase in terrace elevation. The heavy mineral content in the laminations/beds ranges from 80 to 99 percent.

Mineralized sands occur within the supratidal zone of all modern beach tracts on the North Strand but are richest where mineralized terraces are present. Concentrations of heavy minerals are highest in the supratidal zone, at the base of eroding bluffs, and decrease substantially within the intertidal zone. The width of the supratidal zone varies from 10 to 40 m and the zone of mineralization appears to be less than 2 m thick.

The sedimentology of the modern beach sediments is similar to that of the raised terraces. Within the supratidal zone, subhorizontal planar bands of quartzofeldspathic minerals are interlaminated with well-sorted heavy mineral layers (Plate 3). Laminations are gently inclined and thin seaward, and crosscutting erosional surfaces are evident. Heavy mineral layers vary from 0.5 to 40 cm in thickness, and contain from 80 to 99 percent heavy minerals. At Beaver Brook, the heavy mineral assemblage consists of hematite,

Plate 3. *Mineralized beach sands within the supratidal zone of the modern beach.*

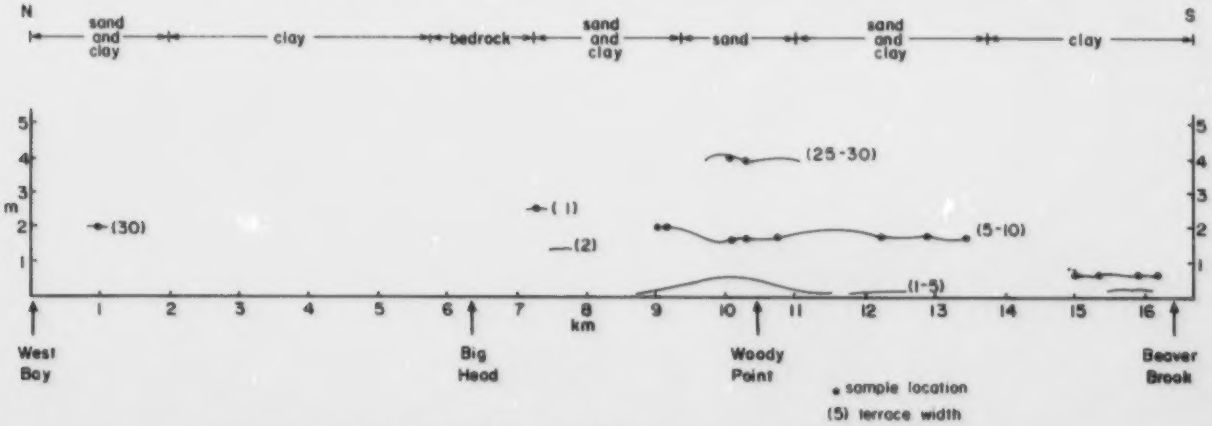

Figure 5. *Distribution of mineralized beach terraces on the North Strand.*

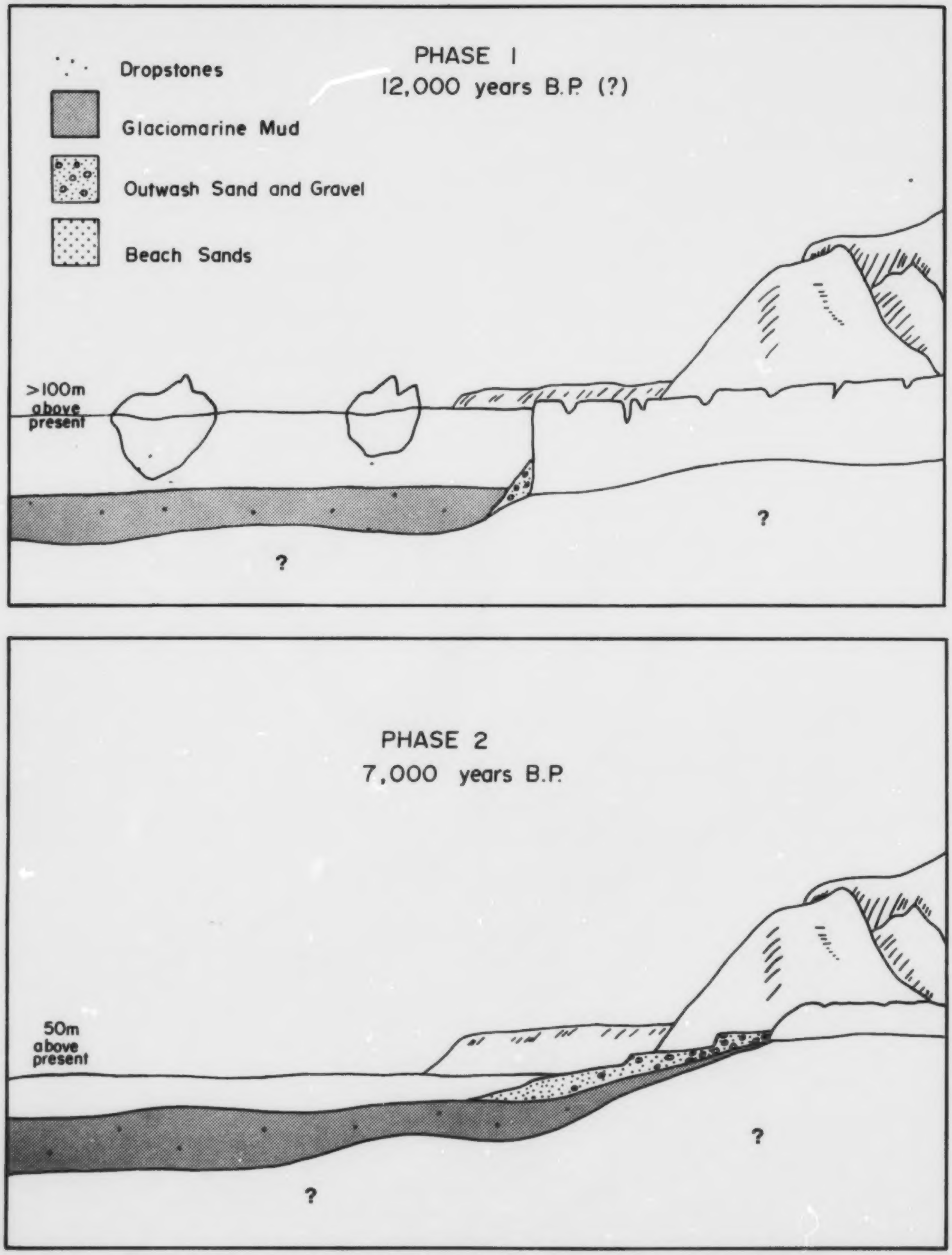
Dropstones
Glaciomarine Mud
Outwash Sand and Gravel
Beach Sands
PHASE 1
12,000 years B.P. (?)
>100m above present
?
?
PHASE 2
7,000 years B.P.
50m above present
?
?

Figure 6. *Three-stage conceptual model depicting coastal evolution and placer formation.*

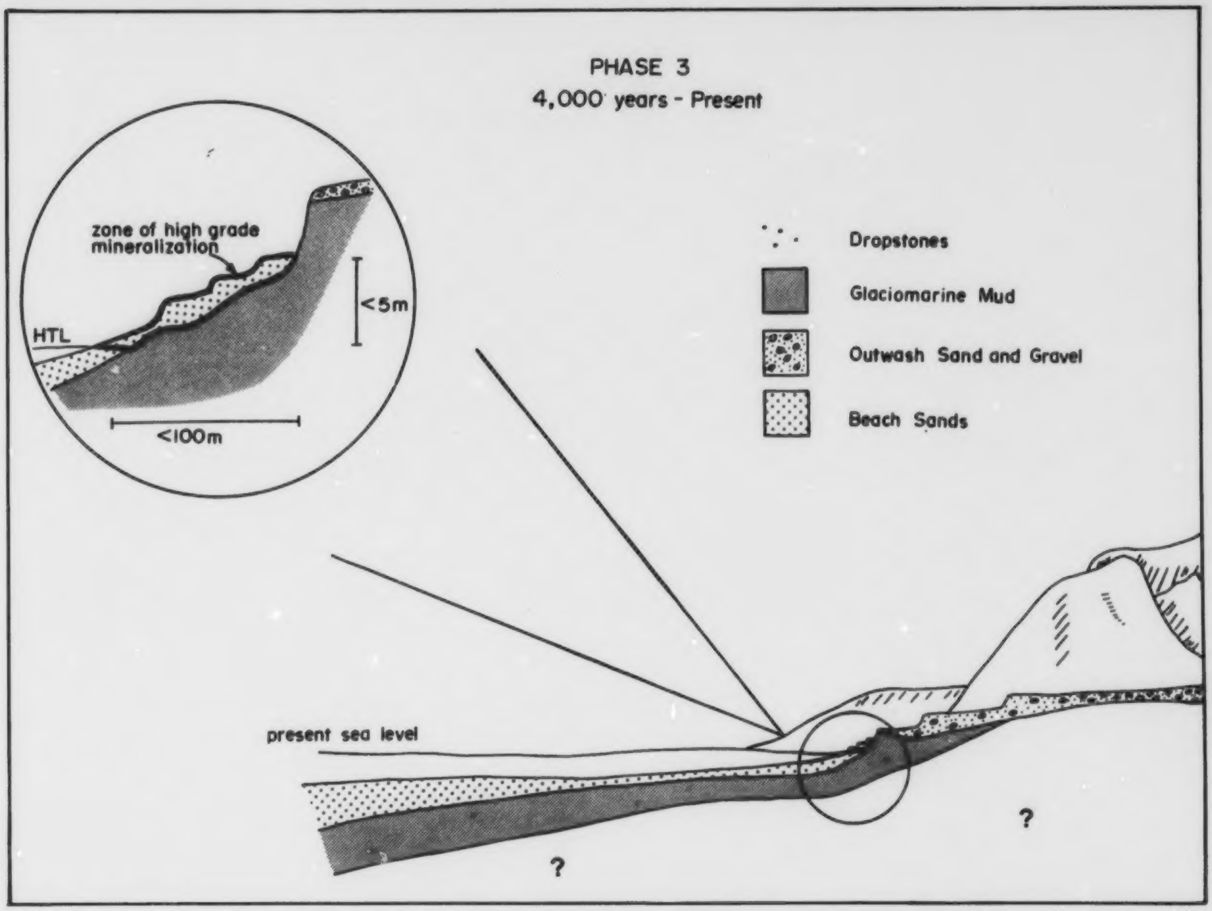

Figure 6. *(Continued).*

amphibole, garnet, magnetite and ilmenite (Mathieu and Boisclair, 1990), which are often found in mono-mineralic laminations. The overall TiO_2 content, determined from vertical channel samples through these sediments, ranges from 1.79 to 7.4 percent, with an average content of 4.6 percent, and total Fe content averages 30.4 percent.

The intertidal sediments exhibit well-defined mineral layering, with the heavy-mineral laminations comprising between 15 and 35 percent of the sampled sections. At Beaver Brook, the intertidal sediments are characterized by an average TiO_2 content of 0.79 percent and a total Fe content of 5.4 percent.

INTERPRETATION AND DISCUSSION

In an attempt to integrate the Quaternary geology of the North Strand, (as outlined by Rogerson, 1977) and the observed distribution of mineralized beach sands, a conceptual model depicting coastal evolution and placer formation is proposed (Figure 6).

Phase 1

At some time during the late Wisconsinan ice recession, (possibly around 12,000 BP; Rogerson, 1977) the ice margin remained stationary for an extended period of time near the present coastal limits. The marine limit was greater than 60 m above present sea level and a thick sequence of glaciomarine mud was deposited. The occurrence of dropstones in the glaciomarine muds suggests that icebergs were present, perhaps calving off the face of a tidewater glacier.

Phase 2

As ice began to recede, substantial volumes of sediment-laden meltwater issued from the front of the ice margin. Sands and gravels were deposited in a large outwash complex that prograded over the glaciomarine mud. The seaward margin of the outwash was deposited within a subaqueous deltaic environment, the stratigraphy of which is revealed in the modern coastal bluffs. During the constructional phases of outwash deposition, relative sea level was dropping at a very

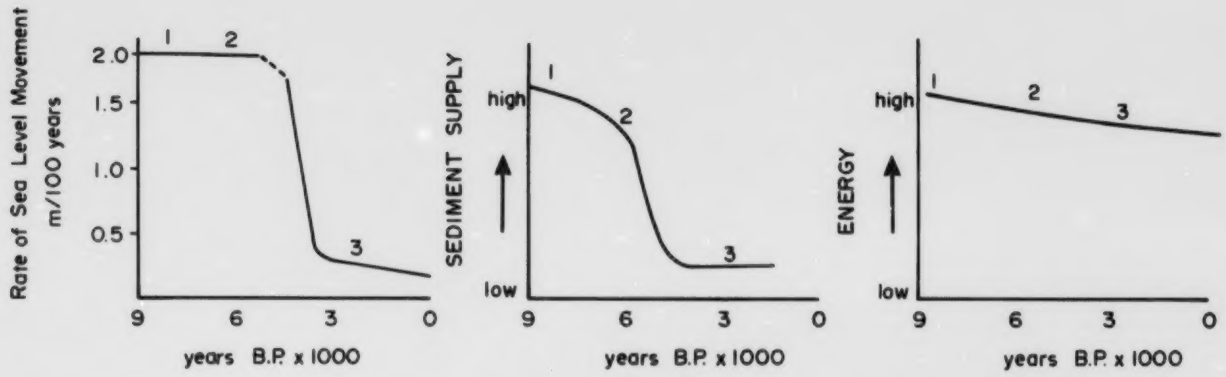

Figure 7. *Temporal variations in the rate of sea-level movement, sediment supply and marine energy.*

rapid rate (1.2 to 2.4 m/100 years). A series of terraces were cut into the outwash sediment during emergence, perhaps as a result of sea-level stillstands or temporary reductions in sediment supply.

The outwash sands and gravels are clearly an important secondary store of heavy minerals but do not hold significant placer potential. Reworking of the outwash sands and gravels on high-elevation terrace surfaces does not appear to have enhanced the mineral potential, possible because of the rapid emergence and the coarse-grained nature of the sediment.

Phase 3

A major change occurred in the depositional setting of the North Strand approximately 5,000 years BP. Ice retreated from the area and meltwater no longer contributed high-sediment loads, the rate of emergence had slowed dramatically (by a factor of 10), and the emergence of rocky headlands resulted in the division of the coastline into discrete beach tracts (e.g., Big Head, Figure 1). As a result of these changes, a period of backshore bluff-cutting and placer formation began.

Mineralized beach terraces were most extensively developed where the coastal bluffs were composed of outwash sand and gravel (e.g., Woody Point). Beaches bordered by clay bluffs (e.g., Beaver Brook) received sands via longshore transport and the resultant terraces, whereas in some areas heavily mineralized, are volumetrically small. Preservation of mineralized terrace sequences is greatest within small embayments at the mouths of brooks and in the lee of offshore islands (e.g., Woody Point).

In summary, the upper non-mineralized terraces formed under conditions of a high-sediment supply (principally through glacial melt), high energy and rapid emergence. The lower elevation mineralized terraces were formed during a period of slow emergence, high energy and limited sediment supply (Figure 7). Clearly, the high energy and the slow rate of emergence on the North Strand have favoured recent placer development but perhaps the low-sediment budget has precluded more widespread mineralization. In some of the large placer districts of the world (e.g., Australia), a sustained supply of sediment is provided through longshore transport of material derived from rivers and/or coastal erosion. During the formation of mineralized terraces on the North Strand, the primary source to the beach complexes was, and continues to be, the erosion of coastal bluffs. Most of the bluffs are, however, dominated by glaciomarine mud (with the exception of Woody Point) and hence the supply of sediment is quite low. Furthermore, the occurrence of rock headlands at several sites along the North Strand constrains longshore sediment transport and perhaps prevents sediment bypassing between beach tracts. A finite volume of beach sediment has thus been continually 'processed' and upgraded, producing very high-grade placers, but of a limited extent.

The 'processing' or dynamics of mineral sorting, which generates the mono-mineralic layering evident in the supratidal sediments of the North Strand, is of considerable interest. It would appear that under storm and high swell conditions, the backshore sediments are eroded and the heavy minerals are selectively sorted through the process of differential grain entrainment. This type of sorting phenomena has been previously examined, (Komar, 1989) but rarely is such a high level of mineral enrichment attained.

ECONOMIC POTENTIAL

Sand samples from testpits in the supratidal and intertidal zones were collected in 1990 from the Beaver Brook area, and sent to CANMET for mineral separation and identification. A composite sample from the supratidal zone has the following concentrations: 27.7 percent hematite, 24.9 percent quartz, 17.1 percent amphibole,, 16.7 percent feldspar, 8.8 percent garnet, 3.9 percent magnetite, and 2.0 percent ilmenite (Mathieu and Boisclair, 1990). However, geochemical analyses of the composite sample was found to have an average TiO_2 content of 4.6 percent, suggesting that the TiO_2 is not entirely found in the free ilmenite. A preliminary investigation of polished thin sections, being carried out at Memorial University, has shown that the TiO_2 is bound in free ilmenite and as ilmenite intergrowths in hematite grains, (R. Patey, personnal communication, 1991).

Placer deposits similar to those of the north Strand have been mined in New Zealand, Philippines, China and Japan (Elsieev, 1980). Iron–titanium placers also occur on the south shore of Quebec, in a geological environment very similar to the north Strand (MacKenzie, 1912). The principal use of titaniferous ores (hard-rock and heavy mineral sands) is in the manufacture of titanium dioxide, a non-toxic, powdered white pigment used in paint, paper, plastics, rubber, and other filler and extender applications (Harben and Bates, 1990). In the manufacture of titanium dioxide, high-grade titaniferous feedstock, usually natural or synthetic rutile, or high-TiO_2 slag is used. One of the world's largest producers of titanium slag is Q.I.T.- Fer et Titane Inc., whose plant at Sorel, Quebec, produces TiO_2 slag from ilmenite ore mined at Lac Allard, Quebec. Q.I.T. is also looking at importing low-alkali ilmenite beach sand from their deposit in Madagascar, for the production of low-alkali slag. There may be opportunities for other producers of competitively priced titaniferous sands with suitable chemistry (i.e., <0.5 percent MgO, and <0.1 percent CaO) (Bergeron, 1990).

SUMMARY

The placer potential of the Porcupine Strand, in eastern Labrador, is being investigated through channel sampling and very shallow drilling. The economic value of these placer occurrences lies in the Fe and Ti content, with garnet sand being a possible economic co-product. Although geochemical results from the 1991 program are not yet available, the greatest potential appears to be in the modern and raised beach sands, where concentrations of heavy minerals are up to 80 percent. Results from the 1990 sampling program indicate that the TiO_2 content ranges up to 7.4 percent.

The coastal evolution and placer formation of the Porcupine Strand has been controlled by the rate of sediment supply and rate of coastal emergence. Initially (12,000 to 5,000 years BP), there was a high volume of sediment supply and rapid rate of emergence, conditions that were not conducive to placer formation. At approximately 5,000 years BP, sediment supply was greatly reduced, and the rate of emergence slowed dramatically, allowing for increased sorting of sediment, and thus placer formation. However, it has been a relatively finite volume of sediment that is continually being reworked, and the resulting placers, although of high grade, may be of limited extent.

ACKNOWLEDGMENTS

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GOLD METALLOGENY OF THE EASTERN DUNNAGE ZONE, CENTRAL NEWFOUNDLAND

D.T.W. Evans
Mineral Deposits Section

ABSTRACT

Epigenetic gold occurrences, hosted by Paleozoic rocks of the eastern Dunnage Zone, are structurally controlled and spatially related to regionally widespread structures. Two broad belts of gold mineralization, an eastern arsenopyrite-rich belt and a western arsenopyrite-poor belt have been identified and may reflect variation in fluid–source area. The gold occurrences are classified as either mesothermal or epithermal. Mesothermal mineralization is subdivided into three subclasses; 1) shear-zone hosted, 2) disseminated, and 3) dilational veins.

Gold mineralization, which is hosted by a wide array of rock types of different ages, is the result of a prolonged Late Silurian–Early Devonian mineralizing event related to Silurian orogenesis. Source rocks for the mineralizing fluids would have been metamorphosed Gander Zone rocks, which were overthrust by Dunnage Zone sequences during the Ordovician and Silurian.

INTRODUCTION

The eastern Dunnage Zone (i.e., south and east of the Red Indian Line) has become the focus of extensive gold exploration since the 1980's. Significant discoveries have been made in the Victoria Lake, Bay d'Espoir, Great Bend, Glenwood and eastern Notre Dame Bay areas (Figure 1). In the fall of 1989, a metallogenic study was initiated in an attempt to classify and document the nature and setting of this mineralization. Work during the 1991 field season concentrated on occurrences in the area extending from Great Bend (Northwest Gander River) to eastern Notre Dame Bay. This work involved detailed mapping, sampling and diamond-drill-core logging.

Access to much of the area is provided by the Trans-Canada Highway and the Gander Bay, Lewisporte and Bay d'Espoir highways and via Route 340 from Lewisporte. Logging roads bisect much of the region and provide good access to the areas north and south of Glenwood, south of Loon Bay and Dog Bay, and east of the Bay d'Espoir Highway. The region is characterized by gently undulating topography comprising heavily forested tracts, large cutover areas, long, narrow lakes and extensive bogs. Thick, glacial till covers much of the region resulting in a paucity of bedrock exposure.

REGIONAL SETTING

Although this study is termed 'gold metallogeny of the eastern Dunnage Zone', the study actually includes all Paleozoic rocks east of the Red Indian Line and west of the Gander Zone *(sensu stricto)*. Rocks of the Dunnage Zone (Williams, 1979), record the development and subsequent destruction of the early Paleozoic Iapetus ocean. The geological evolution of the zone has been divided into two broad temporal stages involving pre- and post-accretionary events (Swinden, 1990):

1) Pre-accretionary volcanism and pre- and syn-accretionary sedimentation in a series of Cambrian to Middle Ordovician island-arcs and back-arc basins: e.g., Gander River Complex (O'Neill, 1991); Great Bend and Pipestone Pond complexes (Kean, 1974); Victoria Lake Group (Kean, 1977); Davidsville Group (Kennedy and McGonigal, 1972) and the Exploits Group, (Helwig, 1969). Emplacement of the Taconic allochthons during the Middle Ordovician resulted in cessation of island-arc volcanism. Closure of Iapetus, during the Late Ordovician and Early Silurian, resulted in the deposition of extensive flyschoid sequences in fault-bound basins in the central and eastern Dunnage Zone, and

2) Post-accretionary events marked by crustal thickening, resulting in Silurian epicontinental-style volcanism and fluviatile sedimentation (Coyle and Strong, 1987); e.g., Botwood Group (Williams, 1962), extensive regional deformation,

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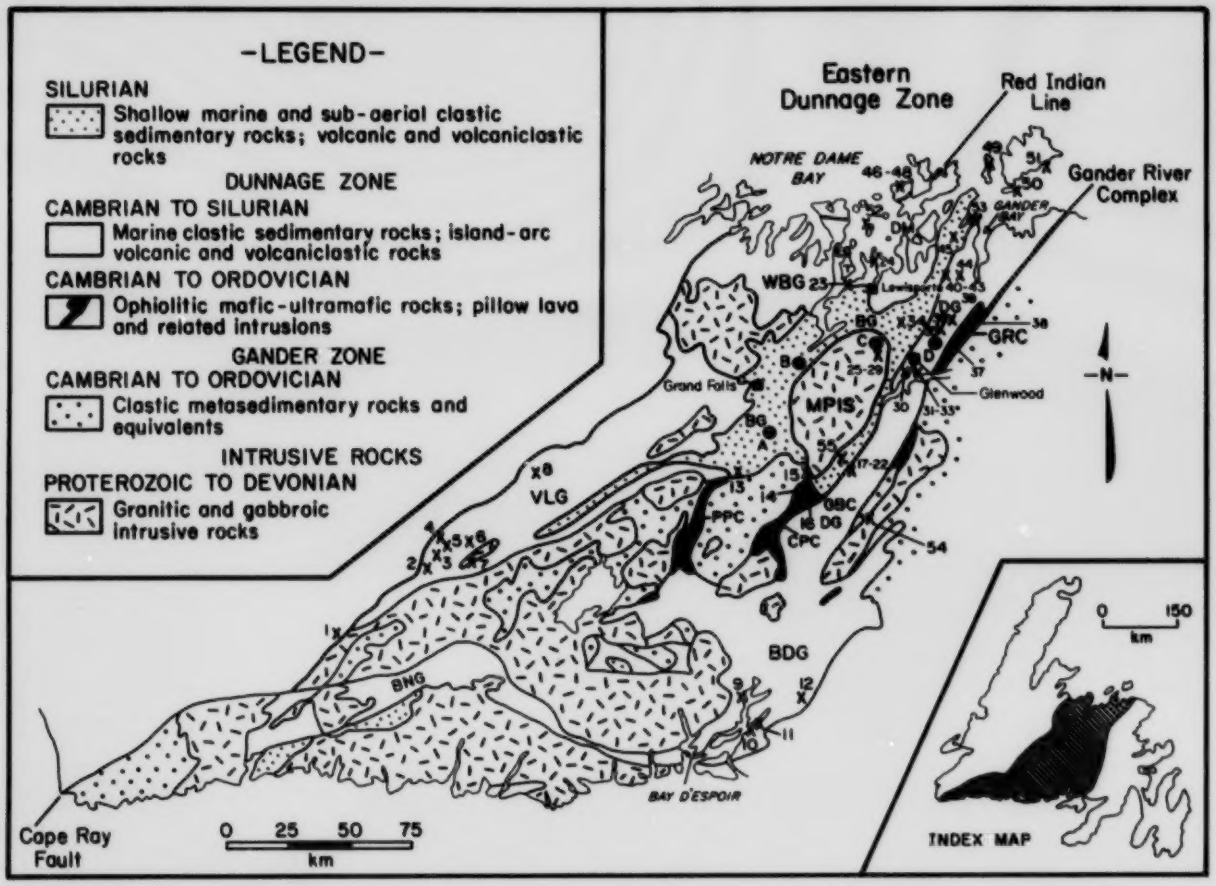
-LEGEND-
SILURIAN
Shallow marine and sub-aerial clastic sedimentary rocks; volcanic and volcaniclastic rocks
DUNNAGE ZONE
CAMBRIAN TO SILURIAN
Marine clastic sedimentary rocks; island-arc volcanic and volcaniclastic rocks
CAMBRIAN TO ORDOVICIAN
Ophiolitic mafic-ultramafic rocks; pillow lava and related intrusions
GANDER ZONE
CAMBRIAN TO ORDOVICIAN
Clastic metasedimentary rocks and equivalents
INTRUSIVE ROCKS
PROTEROZOIC TO DEVONIAN
Granitic and gabbroic intrusive rocks
Eastern Dunnage Zone
Red Indian Line
Gander River Complex
NOTRE DAME BAY
GANDER BAY
WBG
Lewisporte
DG
GRC
BG
MPIS
Grand Falls
Glenwood
VLG
PPC
GBC
CPC
BDG
BNG
BAY D'ESPOIR
Cape Ray Fault
0 25 50 75
km
-N-
INDEX MAP
0 150
km

KEY TO FIGURE 1

	OCCURRENCE	CLASS	GRADES	MINERALOGY	HOST ROCK	ALTERATION
1	SECOND EXPLOITS	Dilational Veins	7.5 g/t Au	(Au)Gn,Sp,Hem,Pyr	Granite	Sil, Epi
2	PATS POND	Dilational Veins	1.9 g/t Au	Au,Ag,Cp,Gn,Sp	Felsic Volc	Sil
3	ROAD (CAMP)	Dilational Veins	5.5 g/t Au	Au,Gn,Sp,Py	Felsic Volc	Seri, Pyr
4	WEST TULKS	Dilational Veins	N/A	Au,Gn,Py	Felsic Volc	Seri, Sil
		Disseminated	N/A	Au,Hem		Sil, Hem
5	MIDAS POND	Shear-Controlled	7.3 g/t Au	Au,Pyr,Tour	Mafic Volc	Fe-carb, Pyr, Intense Argillic
6	LONG LAKE	Dilational Veins	N/A	Au,Pyr	Granite	Sil
7	VALENTINE LAKE	Dilational Veins	24 g/t Au	(Au),Pyr,W,Tour	Trondhjemite, Conglomerate	Seri, Alb, Sil Pyr
8	BOBBYS POND	Epithermal		Pyr,S,Pph,Ser, Alu,Orp	Felsic Volc	Intense Argillic
9	RATTLING BROOK	Dilational Veins	6.5 g/t Au	Au,Bi,Mo	Schist	
10	BOWERS TICKLE	Dilational Veins	13.7 g/t Au	Au,Sb,Ag	Schist	Sil
11	LONG JACKS BIGHT	Disseminated	12.34 g/t Au	Au,Ag,Pyr,Po,Ar,Gn	Schist	Sil
12	LITTLE RIVER	Dilational Veins	10.6 g/t Au	Au,Asp,Pyr,Po,Sb	Felsic Volc	
13	GREAT RATTLING Bk	Shear-Controlled	2.3 g/t Au	Au,Pyr	Ultramafic, Metasediments	Sil, Seri
14	LIZARD POND	Shear-Controlled	12.6 g/t Au	Au,Pyr,Asp	Ultramafic	Sil
15	CHIOUK BROOK	Disseminated	1.9 g/t Au	Au,Pyr,Asp	Altered Seds	Sil
16	BRECCIA POND	Shear-Controlled	< 2 g/t Au	Au	Ultramafic	Sil, Hem
17	AZTEC	Epithermal	<1 g/t Au	Au,Pyr	Altered Seds	Pyr, Argillic
18	HORNET	Dilational Veins	9.7 g/t Au	Au,Pyr,Asp	Granite	Sil
19	A-ZONE EXTENSION	Dilational Veins	2.6 g/t Au	Au,Pyr,Asp	Siltstone	Chlor, Potassic
20	ROAD GABBRO	Dilational Veins	7.9 g/t Au	Au,Pyr,Asp	Gabbro	Sil, Fe-Carb
21	GOOSE	Dilational Veins	1.3 g/t Au	Au,Pyr,Asp	Greywacke	Seri, Sil
22	LBNL	Dilational Veins	1.8 g/t Au	Au,Pyr,Asp	Porphyry	Sil
23	PORTERVILLE	Shear-Controlled	2.12 g/t Au	Au,Pyr,Asp	Gabbro	Fe-Carb Leucoxene
24	POWDERHOUSE COVE	Dilational Veins	78.2 g/t Au	Au,Pyr,Asp	Felsic Dike	Sil

KEY TO FIGURE 1 *(Continued)*

OCCURRENCE	CLASS	GRADES	MINERALOGY	HOST ROCK	ALTERATION
MOUNT PEYTON (25 to 29)					
25 HURRICANE	Shear-Controlled	4.6 g/t Au	Au,Pyr,Asp	Diorite	Seri
26 CORSAIR	Shear-Controlled	3.2 g/t Au	Au,Pyr,Asp	Diorite	Seri
27 COMANCHE	Shear-Controlled	1.3 g/t Au	Au,Pyr,Asp	Diorite	Seri
28 SABRE	Disseminated	2.1 g/t Au	Au,Pyr,Asp	Aplite Dike	Sil
29 APACHE	Shear-Controlled	1.3 g/t Au	Au,Pyr,Asp	Diorite	
30 THE OUTFLOW	Epithermal	12.23 g/t Au	Au,Pyr,Sb	Greywacke	Sil
31 BULLET	Shear-Controlled	83 g/t Au	(Au),Pyr,Asp,Gn,Cp	Shale	Fe-Carb
32 KNOB	Shear-Controlled	155 g/t Au	(Au),Pyr,Asp,Cp,Bou	Greywacke	Fe-Carb
33 BOWATER	Dilational Veins	<3 g/t Au	Au,Pyr	Greywacke	Sil
34 BIG POND	Dilational Veins	440 g/t Au	(Au),Pyr,Asp	Gabbro	Fe-Carb
35 THIRD POND	Dilational Veins	4.6 g/t Au	Au,Py	Greywacke	Sil
36 KNOB HILL	Dilational Veins	2.7 g/t Au	Au,Pyr	Greywacke	Chlor,Pyr
37 JONATHANS POND	Shear-Controlled	6 g/t Au	Au,Pyr,Asp		
38 BURSEY'S HILL	Disseminated	3.5 g/t Au	Pyr (?)	Ultramafic	Talc-Carb
39 CRIPPLE CREEK	Epithermal	9.6 g/t Au	Au,Pyr,Asp	Trondhjemite	Sil
DUDER LAKE (40 to 43)					
40 FLIRT	Dilational Veins	N/A	Au,Pyr,Asp	Gabbro	Fe-Carb, Chlor
41 GOLDSTASH	Disseminated	12.5 g/t Au	Au,Pyr,Asp	Gabbro	Sil,Seri,Fe-Carb, Leucoxene
42 CORVETTE	Disseminated	N/A	Au,Pyr,Asp	Gabbro	Sil,Seri,Fe-Carb, Leucoxene
43 STINGER	Shear-Controlled	N/A	Au,Pyr,Asp	Siltstone	Seri,Fe-Carb
44 BURNT LAKE	Dilational Veins	N/A	Au,Pyr	Greywacke	Sil
45 CLUTHA	Shear-Controlled	N/A	(Au),Pyr,Asp	Gabbro	Fe-Carb, Sil
MORETON'S HARBOUR (46 to 48)					
46 STUCKLESS COVE	Dilational Veins	20.2 g/t Au	Au,Sb,Asp	Felsic Dike	Sil
47 TAYLERS ROOM	Dilational Veins	13.3 g/t Au	Au,Sb,Asp	Felsic Dike	Sil
48 STEWARTS MINE	Dilational Veins	10.9 g/t Au	Au,Asp,Pyr,Sp	Felsic Dike	Sil
49 CHANGE ISLANDS	Dilational Veins	164.1 g/t Au	Au,Pyr,Po,Cp	Felsic Dike	Sil
50 WESTERN INDIAN ISLAND	Dilational Veins	8 g/t Au	Au,Pyr,Asp	Felsic Dike	Sil
51 CANN ISLAND	Shear-Controlled	3.1 g/t Au	Au,Pyr,Cp,Sp	Mafic Volc	Chlor
52 POND ISLAND	Dilational Veins	<1 g/t Au	Au,Cp,Sp,Sb,Bi,Ag, Asp,Tet	Granodiorite	Sil, Seri
53 CHARLES COVE	Dilational Veins	6.2 g/t Au	Au,Pyr,Asp,W,Cp,Mo	Granodiorite	Sil
54 MIDDLE RIDGE	Shear-Controlled	1 g/t Au	Au,Pyr,W	Granite	Seri, Sil
55 HUNAN	Dilational Veins			Greywacke	Seri, Carb
FLOAT					
A PARADISE LAKE	Epithermal			Sandstone(?)	Sil
B MOOSEHEAD	Dilational Veins	N/A	Au,Pyr,Asp	Sandstone(?)	Sil
C SALMON RIVER	Dilational Veins	N/A	Au,Asp,Ag,Sp,Gn,Cp, Pyr,Po	Diorite	Sil
D PANHANDLER	Dilational Veins	161 g/t Au	(Au)	Greywacke(?)	

Figure 1. *Simplified regional geology of the eastern Dunnage Zone, central Newfoundland showing the locations of the significant gold occurrences (see Table 1) and geology, (after Tuach* et al., *1988). Description and tentative classification are given in the accompanying key. DG-Davidsville Group; BG-Botwood Group; DM-Dunnage Melange; EG-Exploits Group; WBG-Wild Bight Group; VLG-Victoria Lake Group; BNG-Bay du Nord Group; BDG-Baie d'Espoir Group; PPC-Pipestone Pond Complex; CPC-Coy Pond Complex; GBC-Great Bend Complex; GRC-Gander River Complex; MPIS-Mount Peyton intrusive suite; T-Thwart Island. Sil-silicification; Epi-epidotization; Seri-sericitization; Pyr-pyritization; Hem-hematitization; carb-carbonate alteration; Alb-albitization; Chlor-chloritization.*

metamorphism, plutonism, e.g., Mount Peyton intrusive suite, (Blackwood, 1981); Charles Cove Pluton (Patrick, 1956); and the Middle Ridge Granite (Strong *et al.*, 1974) and activation or reactivation of major fault systems and thrusting of Dunnage Zone sequences over rocks of the Gander Zone (Colman-Sadd and Swinden, 1984; Keen *et al.*, 1986). These post-accretionary events are interpreted to be the product of a climactic Silurian orogeny (Dunning *et al.*, 1990). A review of earlier work has been completed by Evans (1991).

GOLD MINERALIZATION

This section details some of the significant occurrences examined during this field season and outlines a preliminary classification scheme for gold mineralization within the eastern Dunnage Zone (Figure 2). Dubé (1990) developed a similar classification scheme for gold deposits in western Newfoundland.

Powderhouse Cove

The Powderhouse Cove showing (Figure 1) was discovered by Mr. Quincy Sheppard in 1989. It is best exposed on the eastern shore of Powderhouse Cove, sporadic exposures can be traced westward across the peninsula and it is reported to outcrop on islands in Southwest Bottom. The showing comprises a 2-m-thick, east–west trending, siliceous, aphanitic felsic dyke, which has intruded black graphitic shales of the Dunnage Melange.

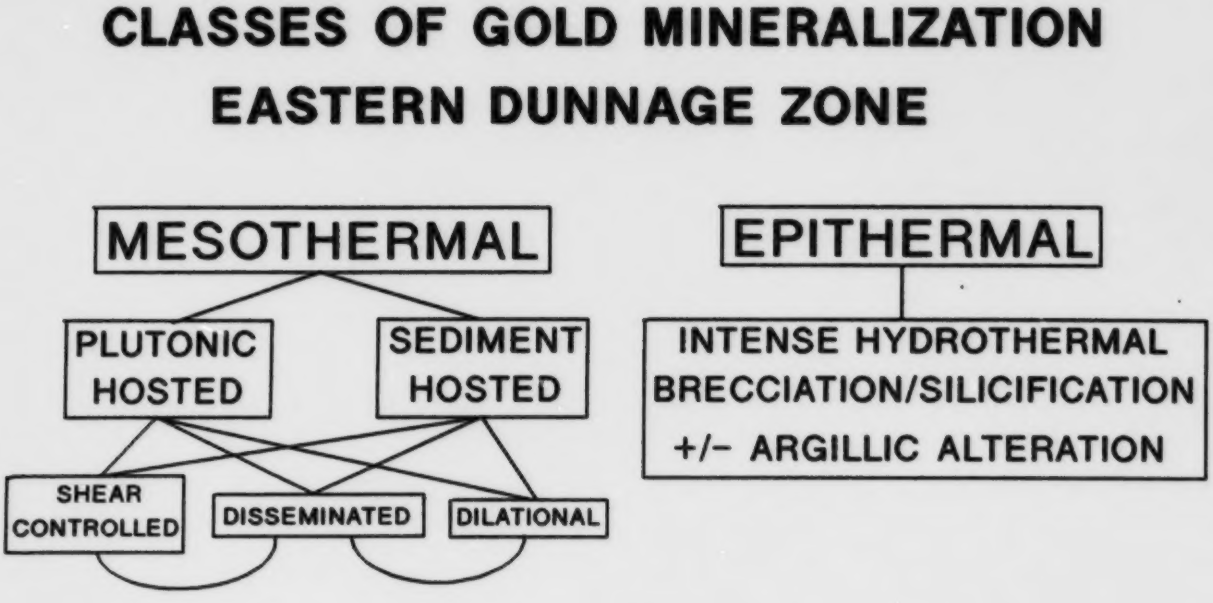

Figure 2. *Preliminary classification scheme for gold mineralization, eastern Dunnage Zone.*

The dyke is interpreted to have intruded during the early stages of regional deformation. This early deformation and intrusion were coeval as is evidenced by a strongly developed east–west-trending foliation that is absent in the shales. Continued deformation produced the main regional cleavage in the surrounding shales and developed a series of foliation-parallel faults that offset the dyke. These faults produced narrow (<3 cm), brittle, quartz-filled fractures within the dyke (Plate 1). Locally, these quartz veins contain massive patches of arsenopyrite and minor pyrite, which have assayed up to 78.5 g/t Au (Table 1). Preliminary results indicate that the mineralization is best developed near the dyke margins, suggesting that carbon complexing may have played a role in the mineralizing process.

Plate 1. *Brittle-controlled auriferous quartz vein (78.2 g/t Au) developed in siliceous felsic dyke, Powderhouse Cove.*

Porterville

Gold mineralization discovered by Mr. Roland Butler in 1990, is associated with narrow shear zones developed within and along the margins of gabbroic intrusive rocks in the Porterville showing area (Figure 1). These rocks, which intrude sedimentary rocks of the Ordovician New Bay Formation, of the Exploits Group (Dean, 1978), are fine to medium grained and locally porphyritic.

The shear zones exhibit Fe-carbonate alteration, minor silicification and local development of leucoxene. Quartz–carbonate veins and stockwork occur within some of the shears and may contain pyrite, arsenopyrite, minor chalcopyrite and sphalerite. The shear zones vary in width from less than 1 m up to approximately 100 m (Butler, 1990). A grab sample from the Porterville showing assayed 2.12 g/t Au. However, assays from the other zones generally assayed less than 2 g/t Au (Butler, 1990).

Similar styles of alteration and mineralization are widely distributed throughout the Bay of Exploits area. O'Brien (1991) stated that gabbro–diorite sheet intrusions in the New Bay area host structurally controlled vein arrays associated with pyritized, Fe-carbonate alteration zones. Similar zones, which assayed approximately 1 percent Cu (Figure 1 and Table 1), were observed on Thwart Island.

Lizard Pond

The Lizard Pond prospect (Figure 1) was discovered by BP Canada Limited in 1989. It comprises a series of sporadic, fault-controlled, quartz–breccia veins (Plate 2) developed within grey-brown, siliceous, locally brecciated magnesite-

Table 1. Selected assays from the eastern Dunnage Zone, refer to Figure 1 for locations. QV-quartz vein; GAB-gabbro; FD-felsic dyke; −not assayed

Sample	Au	Ag	Cu	Zn	Sb	As	Bi	Rock
Thwart Island								
TI-89-03	30 ppb	1.2 g	1.24 %	-	-	-	-	GAB
Charles Cove								
DE-90-181	89 ppb	130 g	-	-	36 ppm	0.57 %	-	QV
DE-89-182	200 ppb	40 g	-	-	33.6 ppm	1.78 %	-	QV
Pond Island								
DE-90-207a	<1 g	1330 g	5.82 %	1.67%	3.67%	1.83 %	0.993 %	QV
Clutha								
DE-89-33e	1.2 g	<2 g	-	-	34.2 ppm	0.78 %	-	QV
DE-89-50e	5.76 g	<2 g	-	-	28.5 ppm	0.04 %	-	QV
Powderhouse Cove								
DE-91-09	92 ppb	-	-	-	-	-	-	FD
DE-91-10a	48 ppb	-	-	-	-	-	-	FD
DE-91-10b	97 ppb	-	-	-	-	-	-	FD
DE-91-10c	145 ppb	-	-	-	-	-	-	QV
DE-91-10d	78.5 g	-	-	-	-	-	-	QV
DE-91-12	222 ppb	-	-	-	-	-	-	FD

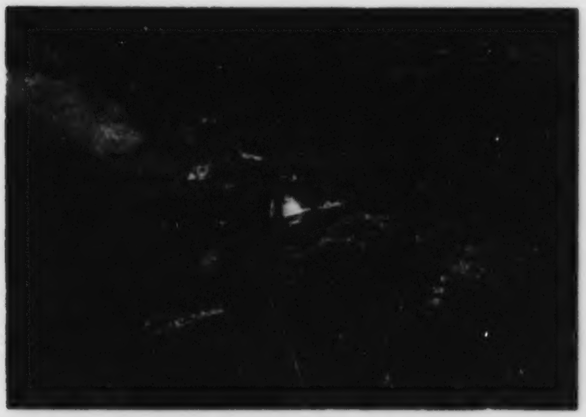

Plate 2. *Structurally controlled quartz−breccia veins developed within silicified magnesite, Lizard Pond.*

altered serpentinite of the Great Bend complex. The veins are up to 1.5 m thick, have strike lengths up to 9 m and contain fine-grained disseminated arsenopyrite and pyrite. Assays from the veins have returned values of up to 12.58 g/t Au over 0.4 m (Graham, 1989).

The altered ultramafic rocks are interpreted, based on limited diamond drilling, to form a series of thin thrust slices separated by red siltstone (Graham, 1989). Based on regional correlations, these siltstones are interpreted to belong to the Silurian Botwood Group (L. Dickson, personal communication, 1991). The structurally controlled gold mineralization within the ultramafic units, if related to thrusting, must therefore be younger than Botwood Group.

Breccia Pond

The Breccia Pond prospect (Figure 1) was discovered by BP Canada Limited in 1989. It comprises a series of structurally controlled, narrow quartz−carbonate veins and minor stockwork having intense hematite alteration haloes (Plate 3). The veins are developed within silicified magnesite-altered ultramafic rock, which contains approximately 5 percent pyrite, arsenopyrite and millerite (Graham, 1989). A channel sample from this prospect assayed 3.24 g/t Au over 1.0 m.

An intensive zone of either hematite alteration or jasperoidal magnesite (Graham, 1989) is developed immediately southeast of the main prospect. Vuggy siliceous zones and small semi-massive patches of pyrite are sporadically developed. Tiny needles of millerite are present in fractures. A diamond-drill intersection revealed that the ultramafic rocks form thin fault-bound slivers separated by shale and sandstone.

Middle Ridge

Auriferous quartz veins, discovered by Noranda in 1988, while following up regional-till and lake-bottom sediment results, occur within the deformed and altered Middle Ridge Granite (Figure 1). The medium- to coarse-grained Middle Ridge Granite, which intrudes sedimentary rocks of the

Plate 3. *Shear-zone-controlled quartz veins with intense hematite alteration haloes developed within altered ultramafic rocks, Breccia Pond.*

Gander Zone, is interpreted as Devonian in age (Bell *et al.*, 1977).

Two generations of quartz veining are recognized (Tallman and Gower, 1988): 1) vuggy quartz veins up to 8 cm wide containing small patches of fine-grained pyrite and minor scheelite assaying up to 1 g/t Au, and 2) 1- to 2-cm-wide cross-cutting, locally anastomosing quartz veins. The vuggy vein set forms siliceous stockwork zones up to 30 cm wide.

Two styles of alteration were also described by Tallman and Gower (1988): 1) brown, pervasive, potassic alteration with disseminated pyrite, and 2) locally developed, narrow zones, 1 to 2 cm thick, of blue silicification containing pyrite and arsenopyrite. The intensity of the potassic alteration is related to both its proximity to the veins and to the degree of shearing within the granite (Tallman and Gower, *op cit.*).

Charles Cove (Tims Harbour)

The Charles Cove showing (Figure 1) discovered by Patrick (1953) comprises an impressive structurally controlled quartz vein developed within the Charles Cove granodiorite (see Plate 11). The showing was prospected, trenched and mapped by NALCO in 1953-54 and Norlex Mines Limited in 1970 for its tungsten potential. This work showed that the north–northwest-striking vein extended for more than 1 km and varied in width from 60 cm to 4.5 m. Three generations of veining were reported: 1) white milky quartz, 2) dark-grey dirty quartz, and 3) white glassy quartz.

A grab sample from the vein assayed 2.81 percent WO_3 (O'Toole, 1967). Grab samples collected by Noranda from the mineralized portion of the vein and containing abundant pyrite, arsenopyrite and minor molybdenite and chalcopyrite assayed up to 520 ppb Au (Green, 1989). A sample containing a 1-cm-wide band of arsenopyrite assayed 6.20 g/t Au. Patches of arsenopyrite up to 30 cm long were reported by O'Toole (1970). The mineralization is reported to be associated with the dark-grey quartz and is restricted to a narrow footwall zone adjacent to the granodiorite.

Green (1989) also reported the presence of additional quartz veins, up to 3 m wide, south of the original showing. These veins contain pyrite, arsenopyrite, trace molybdenite and assayed up to 2.05 g/t Au.

CLASSIFICATION

All of the gold occurrences examined during the course of this study are epigenetic and structurally controlled. Two broad classes or styles of mineralization have been identified: 1) mesothermal, and 2) epithermal (Evans, 1991). Mesothermal mineralization occurs in both plutonic and sedimentary rocks of widely varying ages. This style of mineralization has been divided on the basis of mode of occurrence into three subclasses: 1) shear-zone-hosted auriferous quartz veins; 2) disseminated, and 3) dilational auriferous quartz veins. These subdivisions are not individually exclusive and exhibit some degree of overlap.

Plutonic-hosted mesothermal systems occur in ultramafic, gabbroic and granitic rocks. Ultramafic rocks exhibit two of the mesothermal subclasses: 1) shear-zone-hosted veins; and 2) disseminated. Shear-zone-hosted occurrences, comprising quartz breccia veins (Plate 2), are developed in altered ultramafic rocks of the Great Bend (Lizard Pond and Breccia Pond) and Pipestone Pond (Great Rattling Brook) complexes (Figure 1).

Disseminated gold mineralization hosted by altered ultramafic rocks is restricted to a single occurrence located at Bursey's Hill, Gander Bay Highway (R. Strickland, personal communication, 1991). Gold mineralization (3.5 g/t Au) is hosted by sheared talc–carbonate-altered ultramafic rocks. Further study is required to ascertain the nature of the gold.

Gabbroic rocks are the most prolific host to mesothermal gold mineralization in the eastern Dunnage Zone. All three mesothermal subclasses occur within this rock type with the shear-zone-hosted quartz-vein subclass being the most common. This latter subclass is characterized by shear-zone parallel quartz–carbonate veins (Plate 4) and sigmoidal and oblique vein arrays. The veins are discontinuous, exhibit pinch-and-swell structures and contain patches and disseminations of pyrite, arsenopyrite and locally, free gold. Intense Fe-carbonate alteration haloes, which extend up to 2 m from the quartz veins, are characteristic of this style of mineralization (Plate 4). The best example is developed at the Clutha prospect (Figure 1; Evans, 1991).

The disseminated subclass of gold mineralization within gabbroic rocks has three modes of occurrence: 1) within sheared gabbro, 2) in the wall rock adjacent to the shear zone (Duder Lake; see Churchill and Evans, *this volume*), and 3) in altered wall rock adjacent to quartz veins (Clutha). Intense silicification, Fe-carbonate alteration, disseminated pyrite and arsenopyrite characterize this style of mineralization.

Dilational auriferous quartz veins hosted by gabbroic rocks are similar to the shear-zone-controlled veins in

Plate 4. *Shear-zone-controlled quartz–carbonate vein developed within deformed gabbro, Clutha. Note intense Fe-carbonate halo developed adjacent to the shear.*

mineralogy and wall-rock alteration. The veins exhibit multiple phases of fluid injection and contain abundant fragments of altered wall rock and earlier phases of quartz veining. Wall-rock margins show little evidence for shearing. This deposit type is best exemplified by the Big Pond (Blue Peter) prospect (Plate 5, Figure 1; Evans, 1991).

Plate 5. *Multiple phase dilational quartz–carbonate vein, developed in Fe-carbonate altered gabbro, Big Pond.*

Granitic-hosted auriferous mineralization comprises both shear-zone-hosted and dilational vein subclasses. The shear zones are intensely silicified and sericitized and contain abundant disseminated pyrite, arsenopyrite and quartz stockwork hosted by the Mount Peyton intrusive suite (Figure 1).

Dilational vein systems are developed in granitic or felsic dykes located in the eastern Notre Dame Bay area (Powderhouse Cove and Western Indian Island, Figure 1). These dykes contain brittle-controlled pyrite–arsenopyrite-bearing quartz veins (Plate 1). An impressive dilational quartz vein occurs within the Charles Cove pluton (Figure 1). This vein is the largest observed in the eastern Dunnage Zone and has been previously described in the manuscript.

Sediment-hosted mesothermal systems comprise both shear-zone controlled and dilational quartz vein subclasses. Shear-zone-controlled veins range from less than 1 cm up to 1 m in width (Plates 6 and 7). They comprise shear central, sigmoidal and oblique vein arrays. The veins contain patches of carbonate, pyrite and locally free gold. Fe-carbonate alteration is confined to the shear zone and does not extend into the wall rock beyond the shear. The shear zones vary in width from 30 cm to greater than 10 m. The zones are steeply dipping, oriented at a high angle to the regional foliation and pinch out quickly. The Bullet (Evans, 1991) and Knob (Collins, 1991) prospects (Figure 1) are the best examples of this style of mineralization.

Plate 6. *Shear-zone-controlled auriferous quartz–carbonate veins developed within sheared shales, Bullet. Fe-carbonate alteration is confined to the shear zone.*

Dilational quartz veins developed within sedimentary rocks appear to be restricted to rheologically favourable units such as brittlely deformed greywacke (Bowater prospect, Figure 1, Plate 8). The veins are pyritiferous, vuggy, up to 10 cm wide and locally form narrow quartz breccia zones. Gold assays are typically low (< 3 g/t Au).

Epithermal-style mineralization comprises spectacular hydrothermal breccias and argillic alteration, the best example being the Aztec prospect (Figure 1; Evans, 1991). This prospect comprises a zone of multiple phase brecciation (cockade textures with chalcedonic rhinds) (Plates 9 and 10) and pervasive silicification underlain by a 70-m-thick zone of argillic alteration. Intense hydrothermal brecciation and pervasive silicification are also developed at The Outflow (Evans, 1991) and Cripple Creek prospects (Figure 1; R. Strickland, personal communication, 1991). The Cripple Creek prospect comprises intense multiple-phase hydrothermal brecciation and pervasive silicification developed within trondhjemite of the Gander River Complex.

DISCUSSION

Gold mineralization within the eastern Dunnage Zone forms two broad classes, mesothermal and epithermal. Both

Plate 7. *Shear-zone-controlled auriferous quartz veins developed within sheared greywacke, Knob.*

Plate 8. *Dilational pyritiferous quartz veins developed within greywacke, Bowater.*

classes are structurally controlled and related to regionally extensive structures. The mesothermal systems were relatively shallow and formed either in the brittle-ductile transition zone or in the lower levels of the zone dominated by brittle deformation (Figure 3). Most of the host rocks were metamorphosed in the lower greenschist facies, indicating relatively shallow mineralizing systems. Three mesothermal subclasses are present; 1) shear-zone-hosted vein systems, 2) disseminated, and 3) dialational vein systems. The three subclasses overlap and each exhibits characteristics indicative of the others. Intense Fe-carbonate alteration, silicification, sericitization, pyrite, arsenopyrite and locally leucoxene are characteristic alteration assemblages associated with the mesothermal mineralization. The mesothermal class is independent to rock type; however, the gabbroic rocks appear to possess the best potential for economic mineralization (high grade, low tonnage) because they are rheologically and chemically favourable (i.e., Fe-rich). Similar deposits have been documented from the western Dunnage Zone (Hammerdown and Stog'er Tight; Huard, 1990; Andrews and Huard, 1991), from New Brunswick (Elemtree; Ruitenberg *et al.*, 1990; Tremblay and Dubé, 1991) and from the Archean (Barley and Groves, 1990).

Plate 9. *Cockade-textured hydrothermal breccia, which exhibits multiple phases of brecciation and veining, Aztec.*

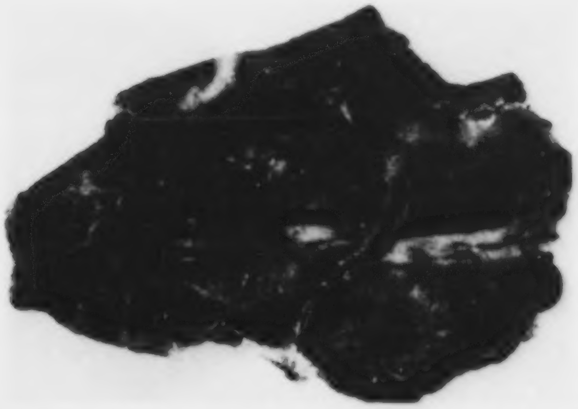

Plate 10. *Multiple phase brecciation, quartz veining and silicification, Aztec.*

The epithermal class exhibits features indicative of shallow or near-surface mineralizing systems (Berger and Eimon, 1983) such as intense multiple-phase hydrothermal breccia (cockade textures and quartz-crystal-lined vuggs), pervasive silicification (silica flooding), argillic alteration and relatively low gold values. This style of mineralization occurs in rocks of the Gander River Complex and the Davidsville and Botwood groups; however, it appears to be more widely developed within the Botwood Group. Intense brecciation occurs north of Big Pond, to the west of the Mount Peyton complex (L. Dickson, personal communication, 1991) and at Great Rattling Brook (MacKenzie, 1990). Float comprised

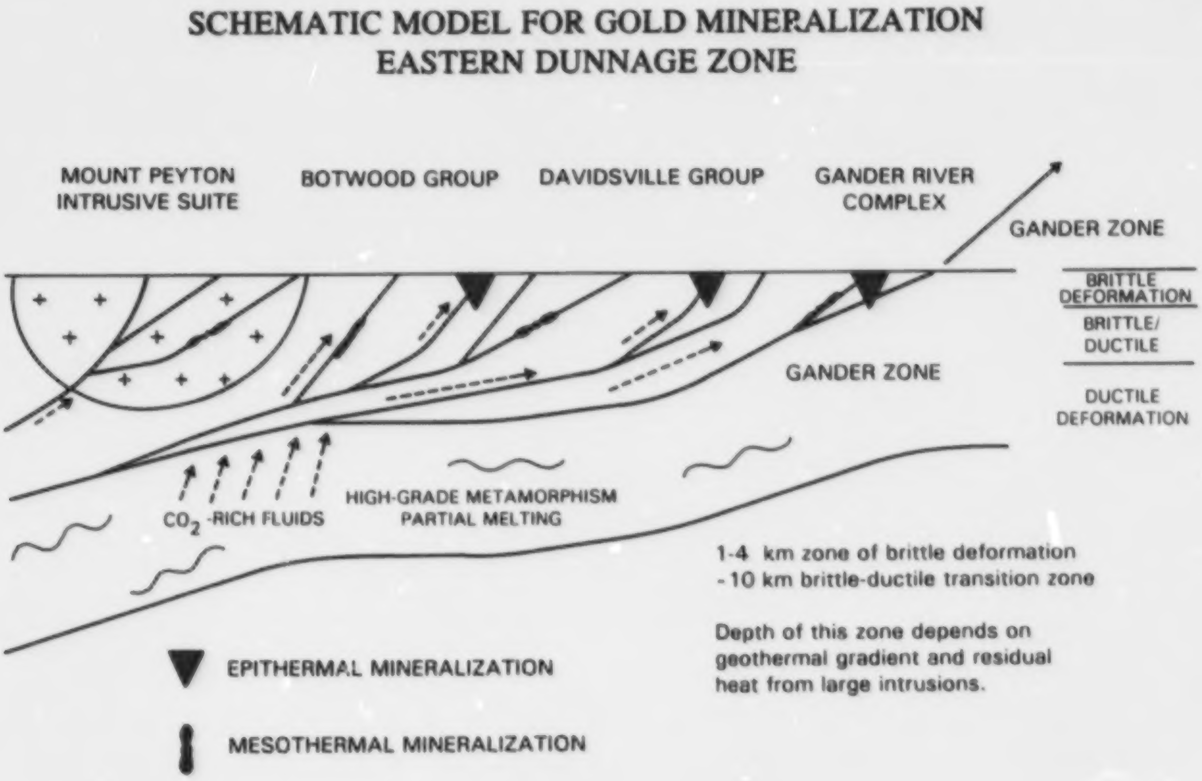
SCHEMATIC MODEL FOR GOLD MINERALIZATION
EASTERN DUNNAGE ZONE
MOUNT PEYTON INTRUSIVE SUITE
BOTWOOD GROUP
DAVIDSVILLE GROUP
GANDER RIVER COMPLEX
GANDER ZONE
BRITTLE DEFORMATION
BRITTLE/ DUCTILE
DUCTILE DEFORMATION
GANDER ZONE
CO2 -RICH FLUIDS
HIGH-GRADE METAMORPHISM PARTIAL MELTING
1-4 km zone of brittle deformation
-10 km brittle-ductile transition zone
Depth of this zone depends on geothermal gradient and residual heat from large intrusions.
EPITHERMAL MINERALIZATION
MESOTHERMAL MINERALIZATION

Figure 3. *Schematic model for gold mineralization in the eastern Dunnage Zone. This model is based on the release of CO_2-rich fluids during progressive metamorphism of Gander Zone rocks. These fluids ascend along thrust faults into second or higher order structures.*

of spectacular hydrothermal breccia containing geyserite eggs was found along the Paradise Lake road, south of Grand Falls (Figure 1). Large angular blocks of this float contain fragments of possible Botwood Group sandstone.

At least two broad belts of structurally controlled gold mineralization can be observed within the eastern Dunnage Zone; a western arsenopyrite-poor belt and an eastern arsenopyrite-rich belt. The western belt is restricted to the southwestern and western portions of the Victoria Lake Group and comprises auriferous quartz veins that contain pyrite, minor base metals and locally tourmaline. Tourmaline is absent from gold occurrences elsewhere in the eastern Dunnage Zone. The eastern belt extends from the Bay d'Espoir area to eastern Notre Dame Bay and includes the eastern portion of the Victoria Lake Group.

Distribution maps of Au, As, and Sb in lake sediments (Davenport and Nolan, 1991a, b; Plate 11) indicate high background levels for these elements, particularly As and Sb, in Exploits Subzone and post-accretionary rocks. These sequences either had high background values to begin with (i.e., are related to syngenetic mineralization) or the element anomalies resulted from epigenetic mineralization related to regionally extensive hydrothermal activity (Davenport and Nolan, 1991a). Both As and Sb are known to be associated with many of the massive sulphide deposits (i.e., massive sulphide deposits within the Tulks Hill volcanics, Victoria Lake Group and absence of arsenopyrite from gold occurrences in the southwestern Victoria Lake Group), suggesting that the distribution may be, in part, primary. However, areas with extensive gold mineralization (Glenwood, Great Bend and Bay d'Espoir) exhibit high concentrations of both these elements in lake sediments suggesting that the As and Sb may be related to gold mineralization.

The widespread nature of both As and Sb, their association with gold mineralization, the higher concentrations of these elements in the Exploits Subzone and post-accretionary rocks together with the spatial relationship between gold mineralization and major regionally extensive structures, favour a widespread, eastern Dunnage Zone, gold-mineralizing event related to late orogenic activity. Dunning *et al.* (1990) proposed that a climactic Silurian orogeny was responsible for the widespread regional deformation, metamorphism, plutonism and subaerial volcanism traditionally attributed to Acadian orogenesis. This orogenesis culminated with the final emplacement of the Dunnage Zone (Exploits Subzone) arc and back-arc allochthon onto the

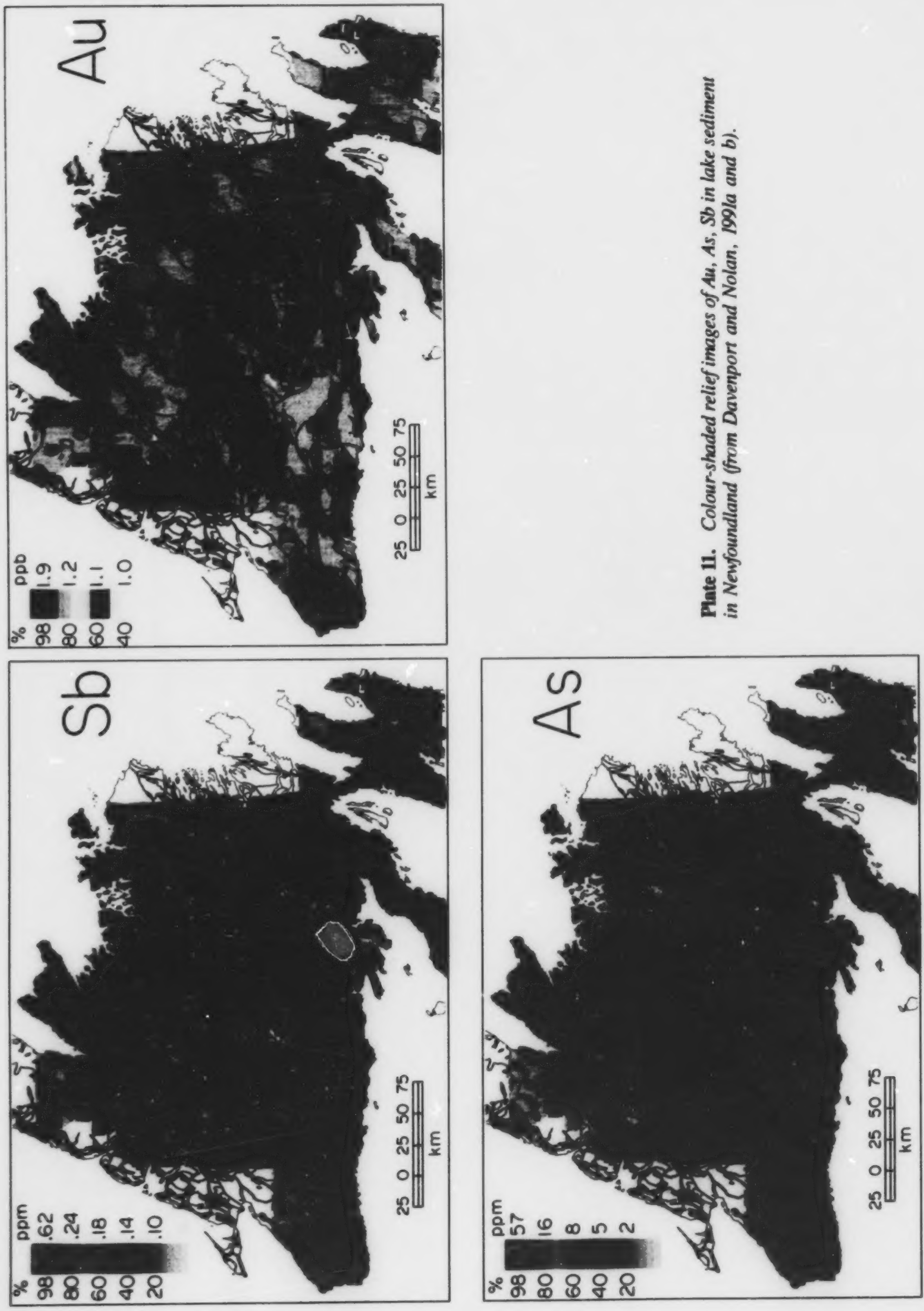
Au
%
98
80
60
40
ppb
1.9
1.2
1.1
1.0
25 0 25 50 75
km
Sb
%
98
80
60
40
20
ppm
.62
.24
.18
.14
.10
25 0 25 50 75
km
As
%
98
80
60
40
20
ppm
57
16
8
5
2
25 0 25 50 75
km

Plate 11. *Colour-shaded relief images of Au, As, Sb in lake sediment in Newfoundland (from Davenport and Nolan, 1991a and b).*

eastern margin of Iapetus (Gander Zone; Stockmal *et al.*, 1987; Williams *et al.*, 1988; Colman-Sadd *et al.*, *in press*). The orogenic event is bracketed between approximately 440 and 390 Ma, as defined by precise U–Pb age dating (Dunning *et al.*, 1990).

The $^{40}Ar/^{39}Ar$ age dating of sericite from an epithermal (Bobby's Pond; Figure 1) alteration zone and from three massive sulphide deposits within the Victoria Lake Group (Kean and Evans, 1988) yielded ages ranging from approximately 405 to 375 Ma, interpreted to represent the thermal metamorphic peak. Similar results ranging from approximately 434 to 374 Ma were obtained by O'Neill (1991) for the age of metamorphism of rocks of the Davidsville Group and Gander Zone.

Gold mineralization is believed to be characteristic of the late stages of convergent orogenic activity worldwide (Robert, 1991) and could be as much as tens of millions of years post-peak metamorphism and plutonism due to the evolution of tectonically overthickened crust. In the eastern Dunnage Zone, this late stage of orogenic activity represented by the thermal metamorphic peak suggests that Dunnage Zone gold-mineralizing processes may have begun as early as approximately 420 Ma (i.e., younger than the Botwood Group and gabbroic rocks that intrude it) and lasted 20 to 40 Ma. Remobilizing of low-grade mineralization along the major structural breaks could have occurred as late as the Carboniferous.

Metamorphic fluids, rich in CO_2 (as evidenced by quartz–carbonate veins, CO_2-rich fluid inclusions and intense Fe-carbonate alteration), could have been derived from the metamorphosed Gander Zone rocks that had been overthrust by the allochthonous Dunnage Zone sequences (Figure 3). The thrusting provided an extensive plumbing system, which allowed the fluids access to the near-surface environments. The nature of the host rocks, i.e., mineralogy (primary and greenschist-facies minerals) and rheological properties (brittle-ductile behaviour) provided conditions favourable for gold deposition (i.e., changes in the physical environment that produced rapid fluctuations in gold solubility and favoured demixing of hydrothermal fluids). Greenschist-facies rocks worldwide are known to provide ideal conditions for the formation of gold deposits (Robert, 1991). Mineralizing fluids responsible for the arsenopyrite-poor belt may have been derived from a different source, possibly the western Dunnage Zone sequence or Grenville basement. Grenville basement is interpreted to underlie a portion of the Dunnage Zone (Keen *et al.*, 1986).

If the metamorphosed clastic sequences (as exposed in the Mount Peyton and Meelpaeg subzones; Williams *et al.*, 1988) were the source of gold and antimony, it may explain why significant mineral occurrences are lacking in these sequences and why they are characterized by low Sb and, locally, Au and As on lake-sediment geochemical maps. Localized high background values of these elements over Gander Zone rocks (i.e., the area north of Gander Lake) may be the product of glacial dispersal. St. Croix and Taylor (1991) indicate that ice-flow direction in this area was from the west, originating in an area with a high density of gold mineralization (Davidsville Group and Gander River Complex).

Future field studies will concentrate on gold occurrences located in the southern portion of the eastern Dunnage Zone, chiefly in the Bay d'Espoir area and on selected occurrences previously visited, thus refining the deposit classification scheme and mineralizing models.

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THE PORT AUX BASQUES COMPLEX: A POTENTIAL SOURCE OF HIGH-ALUMINUM REFRACTORY MINERALS OF THE SILLIMANITE GROUP (KYANITE, SILLIMANITE, ANDALUSITE) AND INDUSTRIAL GARNET

A.F. Howse
Mineral Deposits Section

ABSTRACT

A reconnaissance assessment of industrial minerals in the Port aux Basques region was carried out. The survey showed that gneissic units within the Port aux Basques Complex host significant showings of refractory and abrasive minerals. Schist bands within the gneiss contain kyanite and sillimanite, and in particular concentrations of kyanite in the Grand Bay area. The kyanite occurs mainly in two modes; as relatively large crystal aggregates associated with pods of quartz in gneiss, and as much smaller blades of kyanite, disseminated through and forming part of the schistose fabric of the rock. Garnets are ubiquitous throughout the gneisses and schists, but the best in terms of size, colour, and crystal development, are found in biotite-rich schist bands. The inland region northeast of Grand Bay is a highly prospective area for deposits of disseminated kyanite.

INTRODUCTION

During the 1991 field season, a preliminary assessment of industrial minerals in the Port aux Basques gneiss was carried out. The survey focussed on occurrences of kyanite, a nonmetallic mineral that along with sillimanite and andalusite, share the chemical composition Al_2SiO_5, and are collectively known as the sillimanite group of minerals. All three are industrially important because their product of calcination, mullite ($Al_6Si_2O_{13}$) is highly refractory and in demand wherever high-temperature processes such as smelting of metals and the making of glass are carried on. Refractory applications consume about 90 percent of kyanite production. The remaining 10 percent finds use in numerous products such as ceramic linings, precision castings, and brake shoes and discs. It is also used in the aerospace industry as an insulating material in space capsules and jet aircraft engines.

Garnet is ubiquitous in the Port aux Basques gneiss, and is invariably associated with occurrences of kyanite. It is gaining favour as an abrasive medium because of environmental constraints against its competitors. It contains no free silica, for example, and thus avoids the silicosis problem associated with silica in sand blasting. Also, garnet's specific gravity (3.5 to 4.2) is higher than that of silica sand (2.65), therefore, it requires less air pressure to achieve the same impact in the sanding process. High-quality garnet is used for grinding and lapping glass, ceramics and other materials. Lower quality is used for cleaning and conditioning aluminum and other soft metals. Also, there is a growing nonabrasive use for garnet in water filtration, which requires a material that resists physical and chemical breakdown during filtering and back-flushing (Harben and Bates, 1990). In the United States, about 30 percent of the total garnet consumption is used in water filtration.

PREVIOUS WORK

Previous assessments of industrial rocks and minerals in the Port aux Basques–Isle aux Morts region (Figure 1) have focussed mainly on silica. Gale (1967), during the course of an assessment of pegmatites in the region, collected some chip samples from the quartz vein at Diamond Cove. The analyses revealed a high-silica content and prompted further assessments of the vein as a potential source of flux-grade silica (Butler, 1967; Fleming, 1968). Gale (*op. cit.*) also mentioned that some of the larger pegmatite veins were a potential source of commercial feldspar and further noted that metamorphic rocks of the western part of the region contained abundant kyanite. Prior to Gale's (*op. cit.*) work, Tater (1964) carried out a reconnaissance survey of pegmatite in the region.

Brown (1977), in a report based on his regional mapping of the Port aux Basques Complex noted that schist bands in the Port aux Basques gneiss contained up to 20 percent kyanite. O'Neill (1985) in a M.Sc. thesis study of the Isle aux Morts River zinc prospect described quartz–muscovite schist units containing a high percentage of kyanite. A deposit of garnet at Burnt Island Bay approximately 20 km east of Port aux Basques was mentioned by Snelgrove and Baird (1953) as a potential source of material for the abrasives industry.

Granites of the region were long recognized as potential sources of building stone and some quarry development was carried out around the turn of the century. At Petites, 7 km east of Rose Blanche, 5 small quarries were opened in a red-pink porphyritic granite and some of this material was used as a facing stone on the St. John's courthouse. Blocks of stone from the Rose Blanche granite were also quarried on a small scale for use in lighthouse construction.

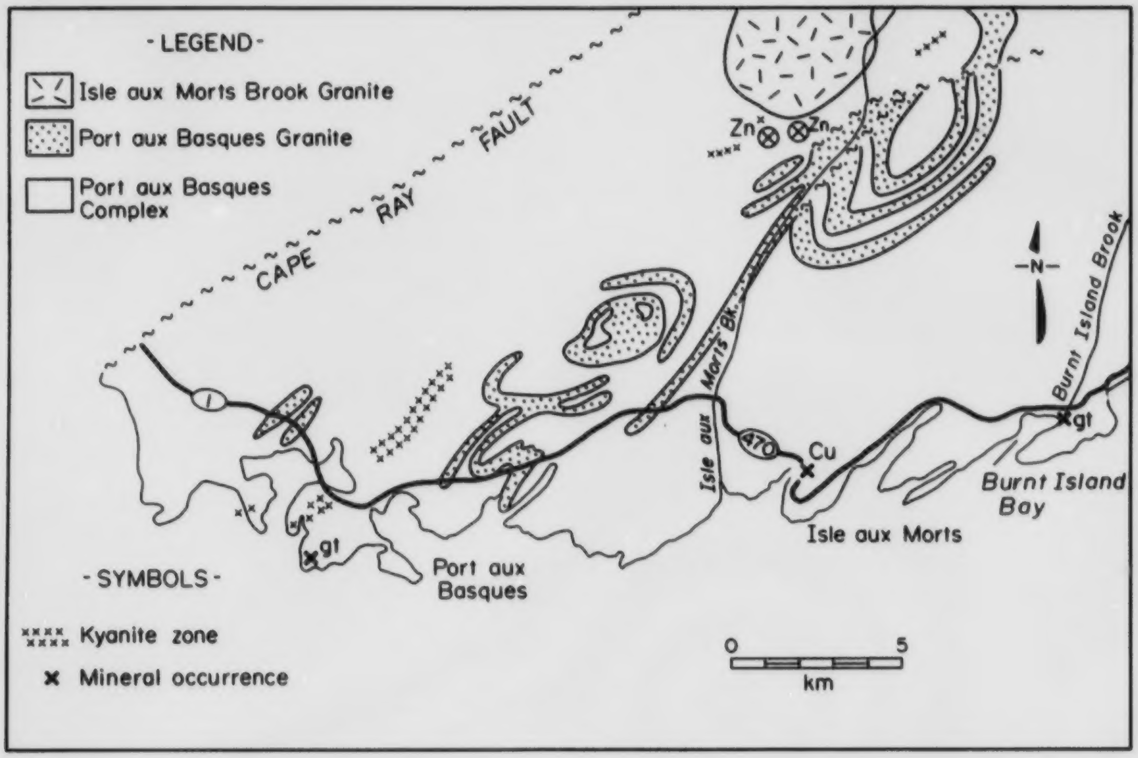
- LEGEND -
Isle aux Morts Brook Granite
Port aux Basques Granite
Port aux Basques Complex
- SYMBOLS -
Kyanite zone
Mineral occurrence
CAPE RAY FAULT
Zn
Zn
Cu
gt
gt
Isle aux Morts Bk.
Burnt Island Brook
Burnt Island Bay
Isle aux Morts
Port aux Basques
470
1
N
0
5
km

Figure 1. *Sketch map showing the regional geology of southwest Newfoundland (after Chorlton* et al., *1984).*

PRESENT SURVEY

OBJECTIVES AND PARAMETERS

The objective of the 1991 survey was to identify deposits of industrial-grade kyanite and garnet in the Port aux Basques gneiss. For the purpose of this report industrial kyanite means material suitable for use in the refractory industry. For garnet, the term industrial mainly implies suitability for use as an abrasive medium in sand-blasting. However, other potential industrial applications are also considered.

ACCESS, PHYSIOGRAPHY, AND METHODS

The Trans-Canada Highway, Route 470, and a network of smaller roads and trails provide good access to the coastal areas. More remote parts can be reached by boat depending on weather and sea conditions. Inland traverses from the coast, beyond the range of a few km, require ATV or, ideally, helicopter support.

Bedrock exposure on the coast is almost continuous except for some sandy beaches west of Grand Bay. Inland, it ranges from poor to good and typically consists of small lichen-covered outcrops along knobby northeast-trending ridges (Plate 1). Low areas and valleys are characterized by growth of stunted spruce trees that make traversing difficult.

Plate 1. *Typical topography in the Port aux Basques region.*

The survey focussed on a 30 km^2 area around and north of Grand Bay, Port aux Basques (Figure 2). The general field method was to locate kyanite-bearing schist bands on the coast and trace them inland along their northeast strike. A significant amount of time was spent in detailed prospecting along individual outcrops. Representative samples of mineral occurrences were collected for analyses and display purposes. All mineral showings were plotted on air photographs and 1:50 000 NTS map sheets.

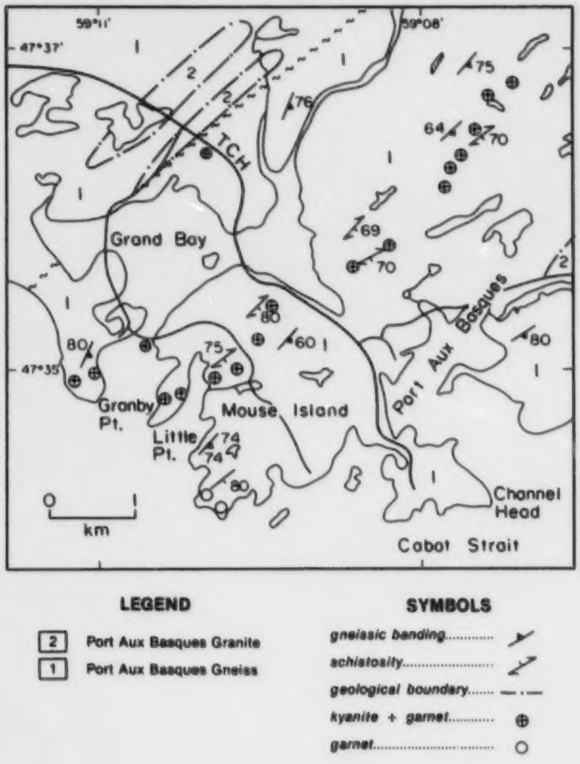

Figure 2. *Kyanite occurrences in the Port aux Basques region; geology after Brown (1977).*

GEOLOGICAL SETTING

The geology of the Port aux Basques region southeast of the Cape Ray Fault is dominated by the Port aux Basques Complex (Figures 1 and 2). As defined by Brown (1972), the Port aux Basques Complex consists of the Port aux Basques gneiss, a well-banded series of light and dark units intruded by granitic rocks, and by pegmatite, aplite, and diorite dykes. The bands, which occur on scales ranging from centimetres to metres, strike northeast and dip steeply to the southeast. The dark units consist of dark green hornblende-rich rock and epidosite. The lighter bands vary in composition from quartz–feldspar to muscovite–biotite garnet schist. Kyanite, sillimanite, staurolite, and garnet are developed in the bands. Schist bands within the gneisses around and northeast of Grand Bay are especially rich in kyanite (up to 20 percent or more) and these were the main focus of the 1991 survey.

Kyanite

A mineral's refractoriness depends mainly on its aluminum content. Theoretically, for the sillimanite group this is 62.93 percent, a figure rarely attained in nature because of impurities such as iron, titanium, and alkalies. Commercial grades of kyanite must contain a minium of 56 percent Al_2O_3 and 42 percent SiO_2 with less than 1 percent acid-soluble Fe_2O_3, 1.2 percent TiO_2 and 0.1 percent each of CaO and MgO (Harben and Bates, 1990).

The rapid depletion of the world's known reserves of massive kyanite has, in recent years, resulted in increased research into methods of recovering kyanite from schists (Varley, 1965). The methods developed involve crushing, milling, magnetic separation, and flotation. In the United States, Kyanite Mining Corporation of Virginia, the sole producer of kyanite in North America, uses flotation to produce a 91 percent concentrate, which is dried and passed through high-density magnetic separators to reduce iron content to less than 1 percent. The kyanite quartzite ore is a medium- to fine-grained rock containing 10 to 30 percent kyanite with quartz muscovite, pyrophyllite, rutile, lazulite, topaz and clay minerals (Johnson, 1990). Other countries that use flotation and magnetic separation techniques are France, Sweden and South Africa.

Garnet

The chief industrial use for garnet is in the abrasive industry, therefore, its physical properties are of prime concern. However, its efficiency for abrasive purposes cannot be determined solely by any standardized series of tests thus behaviour under working conditions is the final criterion (Johnstone, 1984). Hardness should be at least 7.5, and on crushing it should break into irregular fragments having rounded or curved edges. It should be capable of producing a full range of sizes between 8 and 200 mesh. Large crystals that have been badly shattered or weathered are undesirable because they are likely to crumble to dust on crushing.

By far the most important commercial garnet is almandine ($Fe_3Al_2Si_3O_{12}$). Its hardness ranges from 7.5 to more than 8.0 and it conchoidally fractures into sharp chisel-like fragments. This combined with a high resistance to physical and chemical attack make it a superior abrasive. The toughness, fracture and colour of almandine, can be improved by heating to 800°C for about 12 hours. This process is used by several manufacturers in the United States.

It is also desirable that grains of crushed garnet intended for use on paper or cloth should have a high capillary attraction so that the glue will cover them completely and adhere to them when they are attached to the backing medium (Johnstone, 1984).

RESULTS

KYANITE

Several zones of kyanite-enriched mica–garnet schist were identified along a northeast-trending belt that extends across Mouse Island and inland from Grand Bay (Figure 2). On Mouse Island, a line of knobby hills and ridges contains spectacular occurrences of coarsely crystalline kyanite along with schist-hosted disseminated kyanite. It was difficult to gauge the extent of the kyanite mineralization because of the discontinuous nature of the outcrops, but disseminated zones, up to 20 m long and several metres wide, were noted in separate outcrops.

Kyanite mineralization occurs in two distinct modes; as large blue crystals and bladed crystal aggregates up to 10 cm in length, usually wrapped around the rims of quartz pods (Plate 2), and (or) as thin-bladed blue and green crystals (up to 2 cm in length), which along with biotite and muscovite, form the schistose fabric of the rock (Plate 3). Both types of mineralization commonly occur together, especially on the peninsulas that form the east and west sides of Grand Bay. However, inland, northeast of Grand Bay, the disseminated form of kyanite is more prevalent. The following are brief descriptions of some of the best kyanite showings identified to date.

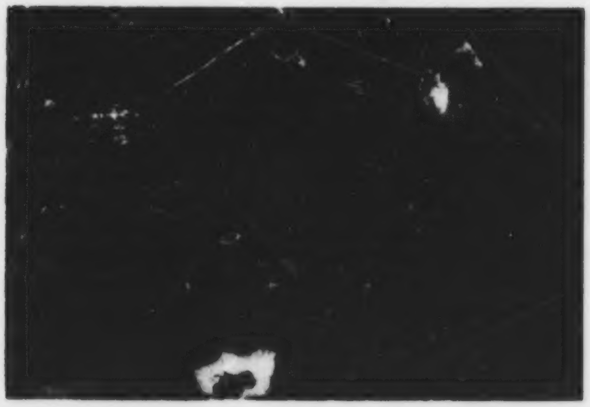

Plate 2. *Coarsely crystalline kyante associated with quartz pods in the Port aux Basques gneiss on Mouse Island.*

Plate 3. *Thin bladed kyanite in kyanite-muscovite-biotite schist bands located inland (2 km) from Grand Bay.*

Inland Showings

A northeast-trending band of kyanite-enriched schist is located about 2 km northeast of Grand Bay and is traceable for several hundred metres along strike. In one of the best occurrences along this zone (Figures 1 and 2), small, 1 to 2 cm blades of blue-green kyanite locally comprise up to 20 percent of the host garnet–mica schist giving the host rock a distinctly blue cast. The hanging wall of this steeply (southeast) dipping zone consists of a sill-like body of highly deformed and crenulated amphibolite schist. The strongest kyanite mineralization is found along a 2-m-wide band bordering this unit although significant amounts are found throughout the length (30 m) and width (8 m) of the mica–garnet schist outcrop. White quartz pods containing books of mica up to 6 cm wide are also present in the kyanite zone. However, this type of mica occurrence appears to be too small for commercial interest.

Grand Bay Showings

Numerous occurrences of kyanite were observed around Grand Bay, notably on the large peninsula on its southeast side, referred to locally (and in this report) as Mouse Island, and on the small peninsula on the southwest side of the bay (Figure 2). One such occurrence is located in a roadcut on the southwest side of the Trans Canada Highway about 500 m east of the Grand Bay intersection. This small showing is the area's most westerly documented occurrence and consists of blue-green crystals up to 4 cm in diameter associated with quartz pods in well-banded gneiss. The showing is too small to be of economic interest.

Granby Point Showing

Several interesting occurrences of kyanite ± staurolite and garnet are exposed in the Granby Point area on the southwest side of Grand Bay. Excellent examples of both coarsely crystalline and disseminated kyanite are present. At both ends of the sandy beach near Granby Point, outcrops of muscovite–biotite–garnet schist host large crystals of kyanite and staurolite. The occurrence at the south end of the beach, the most impressive of the two, features large porphyroblasts of kyanite and staurolite several centimetres in length. A 1-m-wide band of schist contains a high concentration of kyanite and garnet (estimated 20 percent kyanite). Thin biotite-rich schist bands in the rock contain large reddish-brown almandine garnets up to 1.5 cm in diameter.

Similar mineralization can be traced across Granby Point to its western shore about 500 m from the above outcrop. There, in well-exposed northeast-striking, southeast-dipping (80°) units, pale blue acicular crystals of kyanite up to 4 cm in length, are found in crenulated zones of biotite–garnet schist imparting a distinctly blue-green cast to the weathered host rock. A unit of biotite-rich, highly garnetiferous schist occurs near the hanging wall of the kyanite zone. This unit is about 50 cm wide and contains an estimated 40 to 50 percent garnet. Some of the garnets are eroded from their host rock and cover flat areas of the rock.

Causeway Showing

Kyanite occurs in muscovite–biotite–garnet schist at the east end of the causeway connecting Grand Bay East and Grand Bay West. Mottled and crenulated schist contains abundant green acicular crystals up to 2 cm in length. The apparently random orientation of the crystals in the rock gives it a somewhat unique texture. Garnets in the schist are of an uneven size and are deformed and flattened. The showing occurs on a small rock island that now forms part of the causeway and is of no economic significance.

Mouse Island Showings

Several bands of mica–kyanite–garnet schist were identified along a northeast-trending belt that extends across Mouse Island. The showings are best exposed at the crests of low knobby hills and ridges. In some outcrops, the kyanite is quite spectacular, occurring as bladed aggregates up to 10 cm or more and wrapped around the rims of quartz pods. The best examples of this type of kyanite were noted west of Hardy's Arterial Road at the northwest central region of the island. Due to the discontinuous nature of the outcrops, it was difficult to estimate the extent of the kyanite in the schistose bands, although zones up to 20 m strike length and widths up to 5 m were observed. Also noted, especially on the northeastern side of the island, were several outcrops where flat depressions and grooves contained small crystals of eroded kyanite mixed with a sandy matrix (Plate 4). Float, in the form of an angular boulder of banded schist containing disseminated kyanite, was noted about 500 m east of the Hardy's Arterial.

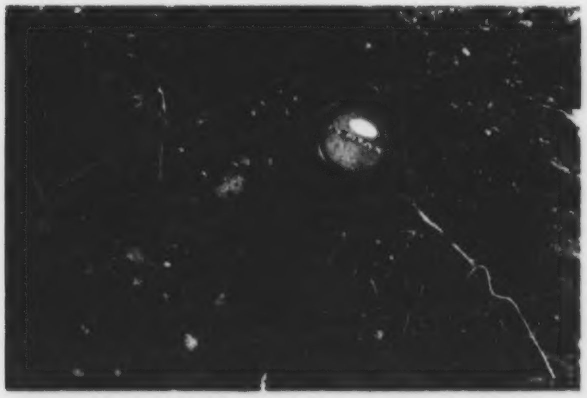

Plate 4. *Small crystals of kyanite eroded from underlying schist band in the Port aux Basques gneiss.*

GARNET

Garnets were observed throughout the Port aux Basques gneiss. Brown (1972) noted an increase in the number of occurrences from the Cape Ray fault toward Port aux Basques with a maximum frequency and size in the Grand Bay area; this observation is supported by the present work. There is considerable variation in the quality of the garnets. The best, in terms of size, colour, and crystal development, were noted in biotite-rich schist bands. In these units, almandine garnets, euhedral to subhedral in outline, comprise up to 40 percent of the rock. However, these units appear to be thin (generally less than 50 cm). Garnet deposits of this type were noted in several outcrops on Mouse Island notably at Granby Point (described earlier) spatially associated with kyanite zones, and in the south cental tip of the island.

Garnets were invariably observed with kyanite in the schist. They are small (usually less than 5 mm), various shades of red and commonly contain other minerals as inclusions. Garnets were also commonly noted at the contact of amphibolite bands with the other rock units and also within the amphibolite (Plate 5). Some of the garnets are up to 1 cm in diameter, are fractured, and contain mineral inclusions.

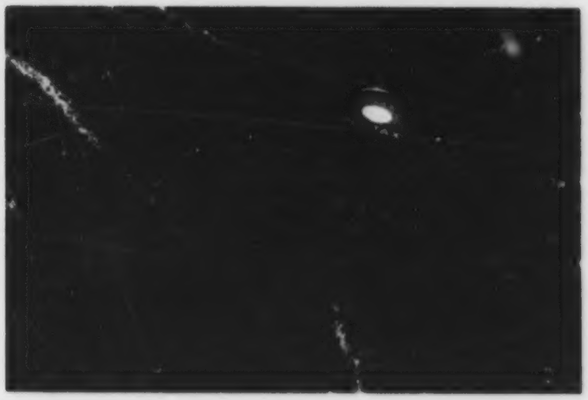

Plate 5. *Garnetiferous amphibolite in the Port aux Basques Complex.*

A garnet showing in Burnt Island Bay, near the mouth of Burnt Island Brook (Figure 1), was investigated during the present survey. The occurrence was mentioned by Snelgrove and Baird (1953), and again by Carr (1958) as a potential source of commercial industrial garnet. The garnet is exposed in concentrated zones in schist bands (050°/80° northwest) and can be traced along a string of small island outcrops that extend offshore for about 400 m. The richer zones range in width up to 2 m and the garnet can be traced more or less continuously through a width of 20 m. The garnets are mottled brown, subhedral and attain diameters up to 1 cm, although the average size is considerably smaller. The showing could not be traced inland because of overburden and the presence of houses due to the expansion of the community of Burnt Island.

SUMMARY

Bands of mica–garnet schist in the Port aux Basques gneiss are good prospective units for high-aluminum refractory minerals of the sillimanite group (mainly kyanite). Locally, kyanite content in the rock ranges up to 20 percent. It occurs mainly in two modes, as large blue-green crystals associated with quartz pods, and as smaller bladed crystals that form part of the schistose fabric of the rock. These crystals can be colourless, light or dark blue or light green, and range up to 2 cm in length. This disseminated type of kyanite mineralization is the most commercially interesting because it is pervasive and can be traced over significant strike lengths. The inland area northeast of Grand Bay is a promising area for further work in light of the showings already identified.

Garnets are ubiquitous in the Port aux Basques gneiss and are abundant in the schistose bands. They are particularly large and well developed in biotite-rich schist bands. In some rock units, the garnets are fractured and contain inclusions of other minerals. This would probably preclude their use in industry.

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PRELIMINARY REPORT OF THE STRATIGRAPHY AND MINERALIZATION OF THE NUIKLAVIK VOLCANIC ROCKS OF THE FLOWERS RIVER IGNEOUS SUITE

R.R. Miller
Mineral Deposits Section

ABSTRACT

The peralkaline granites and volcanic rocks of the Flowers River Igneous Suite have been identified as good targets for rare-metal mineralization. A program of mapping and sampling was initialized in the summer of 1991 to help evaluate the volcanic rocks (Nuiklavik volcanic rocks) of this suite for rare metals (e.g., Y, Zr, REE).

The Nuiklavik volcanic rocks underlie a roughly circular-shaped area 14 km in diameter. Rocks mapped include: feldspar–quartz porphyry, quartz and quartz–mafic porphyry, ignimbritic flows, aphyric and quartz-poor flows, amphibole-bearing ignimbrite and porphyry, and feldspar–quartz–amphibole porphyry. Stratigraphic sections indicate that the preserved volcanic rocks are at least 300 m thick.

Although geochemical results for rare metals are not yet available, radioactivity measurements indicate that the rare-metal content is probably very high for some rock units. Measurements up to 3000 cps (20 times background) were recorded for ignimbrites, aphyric flows and quartz-poor flows. These more radioactive rock units occur in the upper portion of the stratigraphic section where they form mappable units up to 50 m thick. Geochemical and other laboratory studies are on-going.

INTRODUCTION

The Flowers River Igneous Suite (FRIS) is a Mid-Proterozoic felsic volcanic–intrusive complex that occupies a 70 by 50 km area near the coast of Labrador 70 km south of the community of Nain (Figure 1). The complex is dominated by aegirine and arfvedsonite–riebeckite-bearing peralkaline granite plutons. Felsic volcanic rocks, which occupy approximately 4 percent of the complex, consist of feldspar–quartz porphyries, feldspar–quartz–amphibole porphyries, quartz porphyries, phenocryst-poor or aphyric flows, volcanic breccias and ignimbrites.

Rocks of the FRIS were first mapped in 1953 (Beavan, 1954) by BRINEX during a reconnaissance exploration program. The granites and volcanic rocks in the vicinity were further mapped in 1971, at a scale of 1:250 000, in a Geological Survey of Canada regional mapping program (Taylor, 1979). More detailed mapping, at a scale of 1:50 000, was carried out in the period 1977 to 1982 by the Newfoundland Department of Mines and Energy (Hill, 1981, 1982).

The major host rocks of the Flowers River Igneous Suite are the anorthosite–gabbroic and quartz monzonitic intrusive phases of the Nain Plutonic Suite (NPS) (Figure 2). Other host rocks include Lower Proterozoic and Archean gneisses.

Exploration activity in the Flowers River Igneous Suite commenced in 1953 (Beavan, 1954) when the volcanic portion was visited by BRINEX geologists; no significant mineral

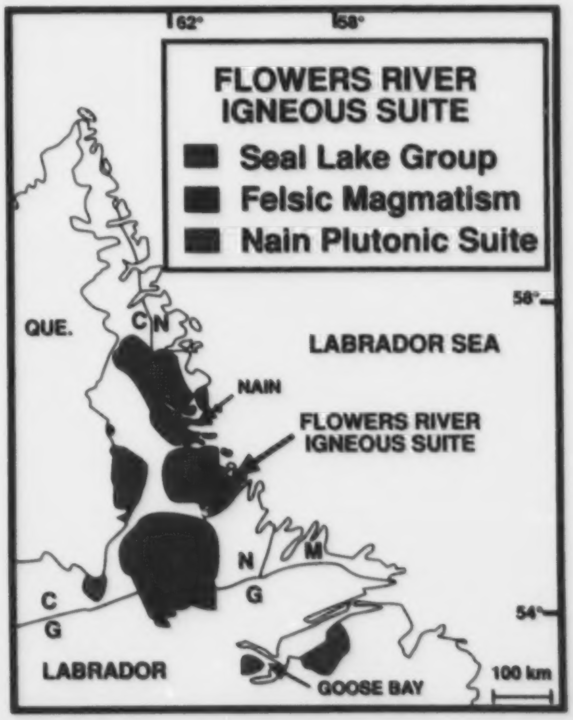


Figure 1. *Location map of the Flowers River Igneous Suite, Labrador. Structural provinces: C = Churchill, N = Nain, G = Grenville, M = Makkovik.*

occurrences were located at that time. A regional lake sediment and water geochemical survey (Geological Survey of Canada, 1978, 1979) carried out under the Uranium Reconnaissance Program indicates that the FRIS, especially the volcanic portion, exhibits anomalies of F, Zn, Mo and Pb. The detailed mapping of Hill (1981, 1982) and the follow-up geochemical work of McConnell (1984) concluded that showings of pyrite, sphalerite, galena, molybdenite and fluorite are common in these volcanic rocks. Hill (*op. cit.*) also noted that high radioactivity is associated with some of the pyrite-bearing volcanic rocks.

The documentation of rare-metal mineralization in the Strange Lake peralkaline granite and the Red Wine Intrusive Suite–Letitia Lake Group resulted in the formulation of an exploration model for rare metals in felsic peralkaline rocks (Miller, 1988). Application of this model to other parts of Labrador indicates that the FRIS is a good rare-metal target. Geochemical analyses of some samples of the FRIS (Hill, 1982; White, 1980), especially the volcanic units, indicate that rare metals, such as Y, Zr and REE (Miller, 1988), reach values that are of economic interest. Consequently, the volcanic rocks were targeted for more detailed mapping and geochemical and petrological study to further evaluate their rare-metal potential. These volcanic rocks were mapped during a 6-week period in the summer of 1991. Laboratory studies on samples collected during this period are on-going.

The volcanic rocks of the Flowers River Igneous Suite were initially called the 'Nuiklavik volcanics' (Beavan, 1954) after the nearby bay, which was called Nuiklavik Dart or Jem Lane Bay on the old British hydrographic charts. Hill (1981) renamed these volcanic rocks and the related granites the Flowers River Igneous Suite after the Flowers River, which cuts through the peralkaline granites and empties into Nuiklavik Dart Bay (later renamed Flowers Bay). In this report, 'Flowers River Igneous Suite' is retained to designate the entire suite, both plutonic and volcanic as Hill had proposed, and the name 'Nuiklavik volcanic rocks', as Beavan had proposed, is retained for the volcanic portion of the FRIS.

This report will outline the results of the 1991 field-mapping program, including a description of the volcanic stratigraphy of the Nuiklavik volcanic rocks of the Flowers River Igneous Suite.

STRATIGRAPHY

The Nuiklavik volcanic rocks were mapped at a scale of 1:50 000. Detailed mapping was carried out in areas of nearly continuous outcrop that were chosen for stratigraphic work (Figure 2). This mapping has identified the following units, which are summarized in the stratigraphic section (Figure 3): upper feldspar–quartz porphyry, upper quartz and quartz–mafic porphyry, ignimbritic flows, aphyric and quartz-poor flows, lower quartz and quartz-mafic porphyry, lower feldspar–quartz porphyry, amphibole-bearing ignimbrite and porphyry, feldspar–quartz–amphibole porphyry and dark feldspar–quartz porphyry. Members of the Nain Plutonic Suite, especially quartz monzonite, form the basement upon which the volcanic rocks were deposited. High-level and medium-grained peralkaline granite intrudes the NPS and some of the Nuiklavik volcanic rocks.

Each unit has been characterized by field observations. Chemical, petrographic and other laboratory studies may result in changes to the stratigraphy presented here (Figure 3).

Nain Plutonic Suite

Rocks of the NPS occur at the contact between the Nuiklavik volcanic rocks (NVR) and the peralkaline granite. They also occur within the outcrop area of the volcanic rocks. Quartz monzonite is the most abundant member, which is associated with subordinate monzonite, diorite and gabbro members. The NPS rocks are crosscut by peralkaline granite and are stratigraphically lower than the NVR.

Peralkaline Granite

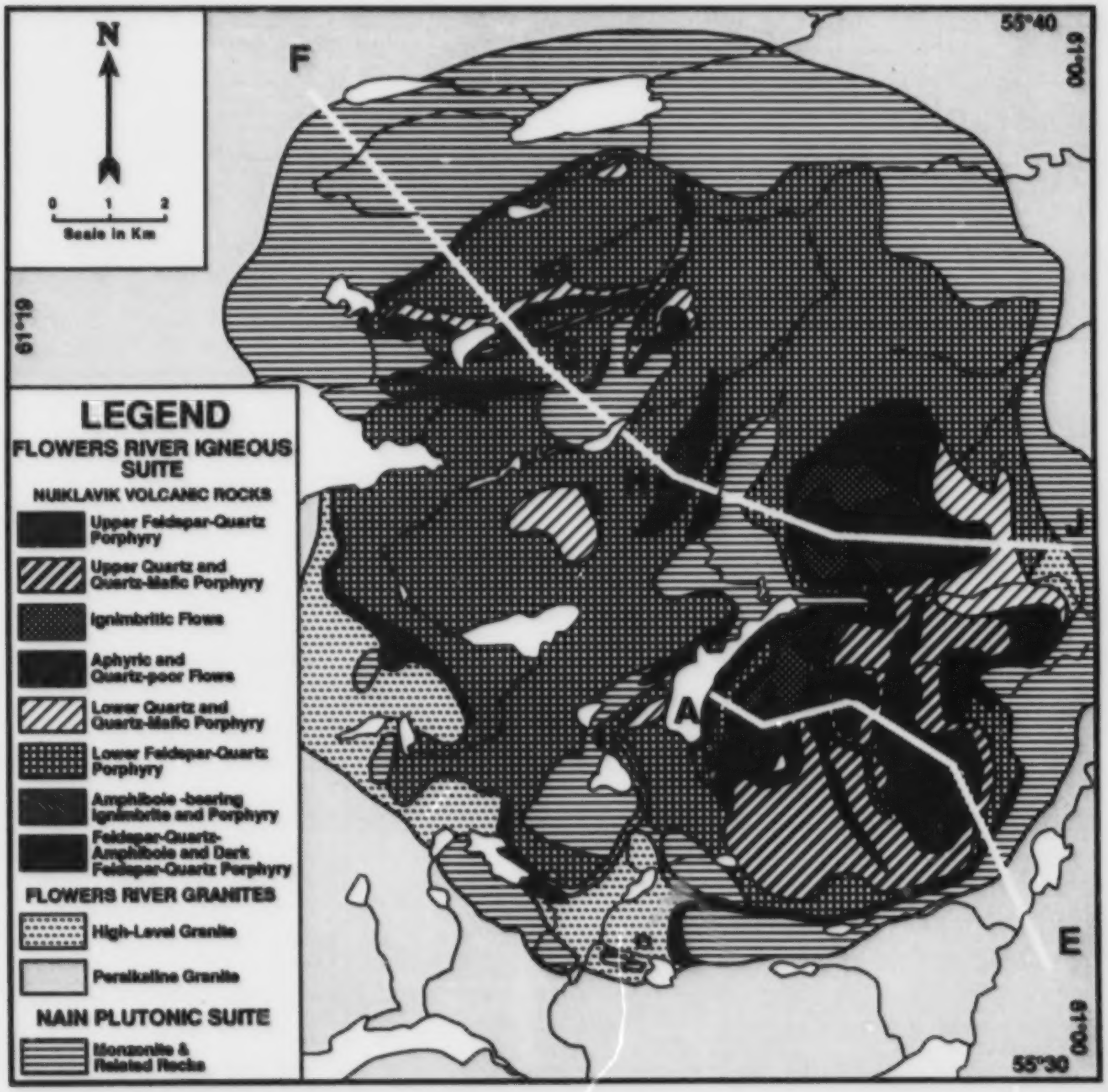
The NVR is completely surrounded by a ring of peralkaline granite, which consists of medium-grained granite, pegmatitic granite and feldspar porphyritic granite. This granite is characterized by riebeckite–arfvedsonite amphibole, quartz and perthitic feldspar. Portions of the granite, particularly in the southwest and western contact zones, are characterized by fine grain-size and rarely aplitic texture; this sub-variety has been mapped separately as 'high-level granite'. Peralkaline granite crosscuts the lower part of the NVR but does not crosscut units above the lower feldspar–quartz porphyry. The peralkaline granite intrudes the NPS.

Dark Feldspar–Quartz Porphyry

This unit is the lowest in the NVR stratigraphy. It is a dark-grey rock that is characterized by up to 20 percent feldspar and quartz phenocrysts that range up to 4 mm in diameter. Many outcrops contain amphibole in the matrix or as small phenocrysts and the matrix is commonly aphanitic to very fine grained. Well-developed columnar joints were observed at one occurrence but more commonly this unit is massive or occurs as rounded to subrounded inclusions in the overlying feldspar–quartz–amphibole porphyry. This unit is at least 20 m thick in some areas but does not appear to be continuous although its occurrence as inclusions in the more wide-spread overlying unit suggests a wider distribution. Field observations suggest that this unit is a volcanic flow.

Feldspar–Quartz–Amphibole Porphyry

This unit is distinguished from the feldspar–quartz porphyry unit above it by the occurrence of up to 2 percent amphibole phenocrysts and fine-grained amphibole in a fine-grained matrix. It commonly contains up to 40 percent quartz and feldspar phenocrysts, which are less than 4 mm in diameter. In some occurrences, it contains inclusions of the dark feldspar–quartz porphyry unit and rarely quartz monzonite of the NPS. It has been observed at or near all


N
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61°00
LEGEND
FLOWERS RIVER IGNEOUS SUITE
NUIKLAVIK VOLCANIC ROCKS
Upper Feldspar-Quartz Porphyry
Upper Quartz and Quartz-Mafic Porphyry
Ignimbritic Flows
Aphyric and Quartz-poor Flows
Lower Quartz and Quartz-Mafic Porphyry
Lower Feldspar-Quartz Porphyry
Amphibole -bearing Ignimbrite and Porphyry
Feldspar-Quartz-Amphibole and Dark Feldspar-Quartz Porphyry
FLOWERS RIVER GRANITES
High-Level Granite
Peralkaline Granite
NAIN PLUTONIC SUITE
Monzonite & Related Rocks

Figure 2. *General geology of the Nuiklavik volcanic rocks of the Flowers River Igneous Suite and the surrounding host rocks.*

observed and most inferred contacts between the NVR and the NPS. It is not clear whether this unit is extrusive or intrusive. Features that contribute to the confusion include: its constant position in the stratigraphy (sill or volcanic flow), its close association with amphibole-bearing ignimbrite (volcanic unit), the abundance of inclusions (derived by intrusion?) and its fine-medium grained texture (similar to the high-level peralkaline granite). These contradictions suggest that this unit may have both intrusive and extrusive phases. The unit is up to 40 m thick.

Amphibole-Bearing Ignimbrite and Porphyry

Several closely related rock types have been grouped into this unit: amphibole ignimbrite, oikocrystic-amphibole porphyry and amphibole porphyry. These rock types are all spatially associated and commonly grade into one another. Features they have in common are the occurrence of up to 30 percent amphibole, few or no quartz phenocrysts and an aphanitic to very fine-grained matrix. Amphibole occurs as single grains, linear aggregates or oikocrystic aggregates. Many samples also contain up to 20 percent red or pink

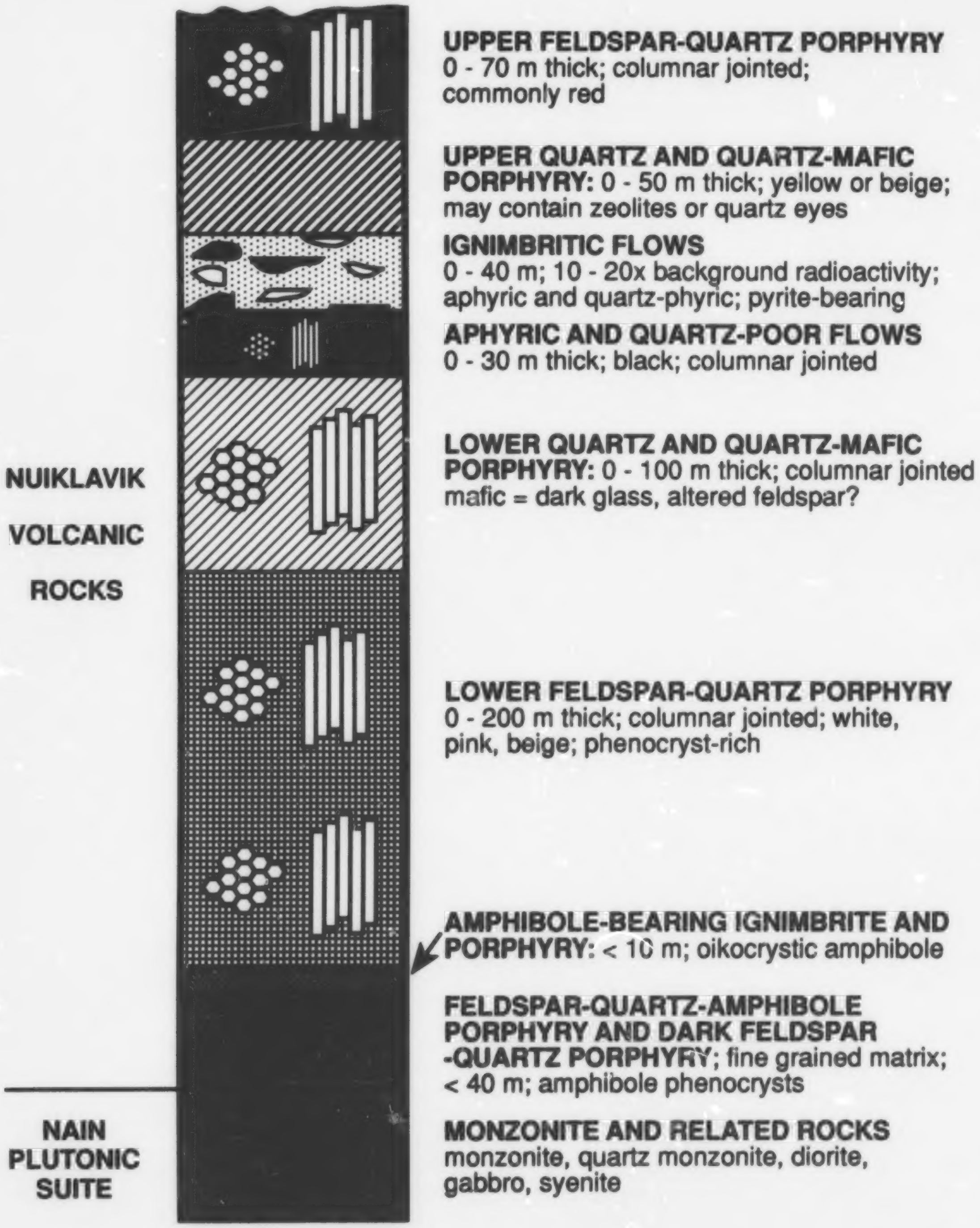
UPPER FELDSPAR-QUARTZ PORPHYRY
0 - 70 m thick; columnar jointed; commonly red
UPPER QUARTZ AND QUARTZ-MAFIC PORPHYRY: 0 - 50 m thick; yellow or beige; may contain zeolites or quartz eyes
IGNIMBRITIC FLOWS
0 - 40 m; 10 - 20x background radioactivity; aphyric and quartz-phyric; pyrite-bearing
APHYRIC AND QUARTZ-POOR FLOWS
0 - 30 m thick; black; columnar jointed
LOWER QUARTZ AND QUARTZ-MAFIC PORPHYRY: 0 - 100 m thick; columnar jointed mafic = dark glass, altered feldspar?
NUIKLAVIK VOLCANIC ROCKS
LOWER FELDSPAR-QUARTZ PORPHYRY
0 - 200 m thick; columnar jointed; white, pink, beige; phenocryst-rich
AMPHIBOLE-BEARING IGNIMBRITE AND PORPHYRY: < 10 m; oikocrystic amphibole
FELDSPAR-QUARTZ-AMPHIBOLE PORPHYRY AND DARK FELDSPAR -QUARTZ PORPHYRY; fine grained matrix; < 40 m; amphibole phenocrysts
NAIN PLUTONIC SUITE
MONZONITE AND RELATED ROCKS
monzonite, quartz monzonite, diorite, gabbro, syenite

Figure 3. *Generalized stratigraphic section of the Nuiklavik volcanic rocks.*

feldspar(?) aggregates. The outcrop area of this unit is very small, as it is only found in seven locations, and it has a maximum thickness of 10 m. The three textural types of this unit may represent at least two different modes of formation: the ignimbritic texture as a gas-charged pyroclastic flow, and the porphyry intrusive dykes or extrusive flows.

Lower Feldspar–Quartz Porphyry

This porphyry is the most widespread of the units in the Nuiklavik volcanic rocks. The thickness ranges from 70 m in the southeast, to 200 m in the northwest. It is characterized by 2 to 10 percent quartz phenocrysts, 10 to 40 percent feldspar phenocrysts, an aphanitic to very fine-grained matrix and phenocryst size up to 5 mm across. Amphibole has been found in a few outcrops. Well-defined, five- or six-sided columnar joints, from 30 to 70 cm wide and up to 10 m long, are very common in larger outcrops of this unit. This texture and the scarcity of matrix textures and fragments indicate that this porphyry was deposited as lava flows or domes.

Lower Quartz and Quartz–Mafic Porphyry

Quartz–mafic porphyry is the most prominent rock type of this unit. It is characterized by 5 to 15 percent quartz phenocrysts, up to 3 mm in diameter, aphanitic to very fine-grained matrix and 5 to 15 percent black to green mafic grains up to 10 mm in diameter. In many cases, the nature of the mafic grains is uncertain in hand-sample but some possible explanations include: altered feldspar phenocrysts, altered glass shards, aphanitic mafic rock fragments and altered zeolitic amygdules. The presence of columnar joints in several outcrops of this unit indicates that it is a flow or a dome. Quartz porphyry, which contains no mafic grains but similar quartz contents, occurs as small volume flows closely associated with the quartz–mafic porphyries. Outcrops of the lower quartz and quartz–mafic porphyries are found throughout the outcrop area. The unit has a maximum thickness of approximately 100 m.

Aphyric and Quartz-Poor Flows

The aphyric or phenocryst-poor flows are characterized by less than 5 percent quartz phenocrysts, which are less than 2 mm in diameter, are black or dark-grey and have an aphanitic to very fine-grained matrix. Some outcrops of this unit also contain up to 8 percent amphibole or mafic spots (mineralogy unknown) and up to 1 percent pyrite. Well-developed columnar joints, which range in size from 10–40 cm across, are also common. Although this unit has flow textures in most outcrops, these textures appear to grade laterally and upward into ignimbritic textures in others. These flows are up to 30 m thick and are found in the northwest and southeast portions of the NVR.

Ignimbritic Flows

Ignimbritic flows in the NVR contain many different kinds of ignimbritic textures but they are characterized by an aphanitic matrix and up to 5 percent quartz phenocrysts, less than 2 mm in diameter. Many ignimbritic outcrops contain aphanitic aphyric fragments and collapsed pumice fragments or fiamme. Ignimbrites are most abundant in the southeastern portion of the NVR and occur as a unit up to 40 m thick.

Upper Quartz and Quartz–Mafic Porphyry

This unit is dominated by quartz porphyry and appears to only occur in the southeastern portion of the NVR. It is a yellow to beige rock characterized by 8 to 15 percent quartz phenocrysts ≤2 mm in diameter, and an aphanitic to very fine-grained matrix. Zeolites and other minerals fill lithophysae in many outcrops indicating that this is a flow unit. In some outcrops, the underlying quartz-bearing ignimbritic flows appear to grade into quartz porphyritic flows. Unit thicknesses range up to 50 m in uneroded sections. In some areas, the lower and upper quartz and quartz–mafic porphyry units form one combined unit as the intervening ignimbritic unit and aphyric and quartz-poor flow unit were not observed; this may be due to poor outcrop exposure.

Upper Feldspar–Quartz Porphyry

This unit is the highest of the NVR units. It is a dark-red rock, characterized by 10 to 45 percent feldspar phenocrysts, 10 to 15 percent quartz phenocrysts, and has an aphanitic matrix. It is found on the tops of the highlands in the southeastern part of the NVR where its thicknesses ranges up to 70 m. Well-developed columnar joints are common.

STRUCTURE

The structure of the Nuiklavik volcanic rocks is illustrated in Figure 4 (cross-section A–E) and Figure 5 (cross-section F–J). Dips of contacts and flow banding range from 20° to 45° toward a centre, which is found in the southeastern or thicker part of the structure. Dips of volcanic units at or near the outer contact with the Nain Plutonic Suite are near vertical. These observations suggest that the structure forms a saucer-shaped volcanic pile having a steep lip. Topographic lineaments, slickensides and the stratigraphy indicate that the volcanic pile is cut by many steeply dipping normal faults, some of which are illustrated in Figures 4 and 5. The obvious complexity of this fault system and the lack of outcrop make it difficult to place faults on the plan view of the volcanic suite.

ECONOMIC GEOLOGY

Several indications of mineralization, in addition to those outlined by Hill (1981, 1982), were located during this field program. Base-metal and radioactive showings are common.

Many outcrops of the ignimbrite flow unit, aphyric and quartz-poor flow units and the quartz porphyry unit contain veinlets or disseminated grains of pyrite and other sulphides. Sphalerite, in association with pyrite, was identified in several flows of the aphyric unit. Galena was observed in a peralkaline granite dyke cutting the NVR. All sulphide showings are minor in extent and rarely contain more than 1 percent pyrite and other sulphide minerals.

The Nuiklavik volcanics, especially the quartz-poor and ignimbritic flow units, contain a large number of radioactive showings that are greater than 5 times background values (i.e., greater than 750 cps). Most amphibole-phyric and quartz-phyric flow units and some aphyric flow units have radioactivity values over 600 cps with selected outcrops up to 2000 cps. All outcrops of the ignimbrite unit have radioactivity measurements of at least 500 cps, with many

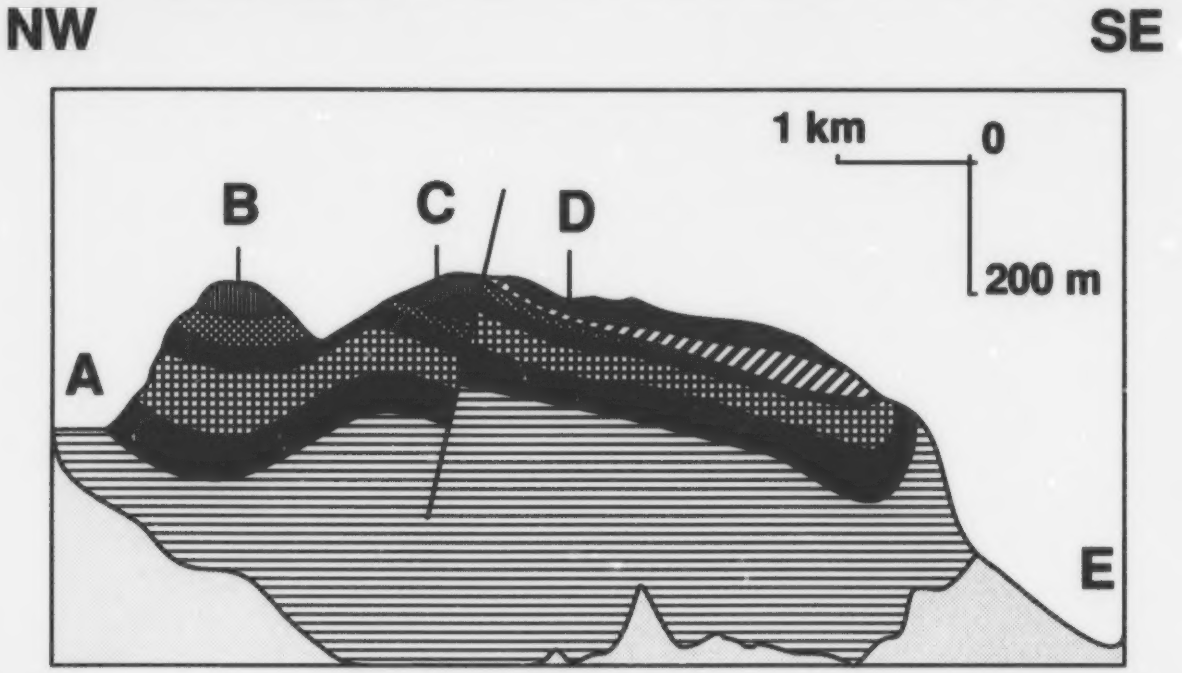

Figure 4. *Cross-section A to E through the Nuiklavik volcanic rocks from the centre of the outcrop area toward the southeast across the outer contact. See Figure 2 for cross-section location and legend.*

in the range 800 to 2000 cps; one radioactive showing in quartz-phyric ignimbrite has values around 3000 cps. Most feldspar–quartz porphyries, quartz–mafic porphyries, feldspar–quartz–amphibole porphyries and peralkaline granites, and many aphyric flows measure less than 400 cps. One peralkaline granite aplite dyke, with textures similar to aplite at Strange Lake, has a value of 940 cps.

Radioactive values are high in rare-metal occurrences in other peralkaline felsic suites in Labrador (e.g., Strange Lake, Miller 1986; Letitia Lake, Miller 1987). The abundance of radioactive showings with high values, similar to those at Strange Lake and Letitia Lake is encouraging for rare-metal exploration.

SUMMARY DISCUSSION

The Nuiklavik volcanics were deposited onto subaerially exposed igneous rocks of the Nain Plutonic Suite. The contact between the volcanic rocks and their basement indicates that the Nuiklavik volcanic rocks were deposited upon a highland during a period of rapid uplift; red beds or conglomerate sediments are absent at this contact.

Amphibole-bearing porphyries and ignimbrites are the oldest exposed volcanic units. The common presence of peralkaline amphibole in these units indicates that at least the lower portion of the NVR is peralkaline in the classical sense. Some of these porphyries may be near-surface intrusive sills or dykes.

The lower quartz and feldspar–quartz porphyry units account for between one third and two thirds of the preserved NVR. The feldspar–quartz porphyry unit is much thicker and much more dominant in the northern and western portions of the NVR. These porphyritic units are not as thick in the southeastern portion of the NVR with the quartz porphyry unit being more dominant there than the feldspar–quartz porphyry unit. These porphyries appear to have originated from several local vents or domes, which are marked by breccia zones.

Ignimbrite and lava flows are very closely associated both texturally and spatially. Both have aphyric and quartz-poor phases, they are dark-coloured and commonly contain pyrite or iron staining. In some outcrops, where they appear to grade into one another, it is difficult to distinguish a highly welded ignimbrite from a banded flow. Aphyric-flow rock fragments are commonly found along with the ubiquitous collapsed pumice fragments in ignimbritic flows. Columnar jointed and massive flows in some outcrops contain peralkaline amphibole phenocrysts and are thus peralkaline in the classical sense.

The upper quartz, quartz–mafic and feldspar–quartz porphyry units are only found in the southeastern portion of the NVR where they form the tops of hills and make up to one third of the stratigraphic thickness. The feldspar–quartz porphyry is the youngest preserved unit in the NVR.

Faulting and subsequent erosion has exposed the underlying NPS in some places. Some of these occurrences of NPS occupy valleys formed by erosion, but others occur

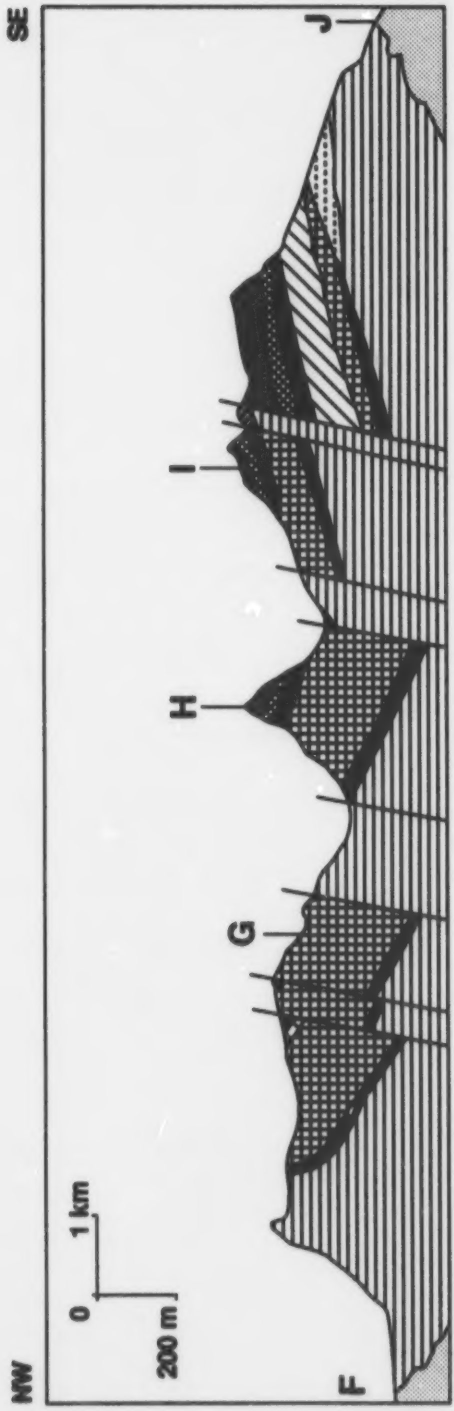

Figure 5. *Cross-section F to J through the Nuiklavik volcanic rocks from the northwest contact through the volcanic rocks and across the southeast contact. See Figure 2 for cross-section location and legend.*

on the tops of hills that are part of upward thrust fault blocks. The timing of the faulting episode is unknown although the interpretation of Hill (1991) suggests that the faulting is due to cauldron subsidence during the extrusion of the NVR.

The intrusion of peralkaline granite occurred during and/or before the extrusion of the NVR. Sills and dykes have been mapped crosscutting the earliest NVR units, including the lower feldspar–quartz porphyry unit.

Field evidence suggests the locations of at least four of the vents that produced the NVR. One vent in the northern portion of the NVR outcrop area is characterized by a breccia of aphyric and feldspar–quartz porphyritic fragments up to 1 m in size. A second vent is located in the northeastern portion of the outcrop area and is identified by breccia fragments of feldspar–quartz porphyry, quartz porphyry and ignimbrite flow. Two vent locations are also tentatively located in the eastern and southeastern portions of the NVR. One is identified by a thick ignimbritic unit (30 to 50 m thick) and an area of extensive quartz-vein stockwork. The other is suggested by thick (up to 35 m) rock-fragment-rich ignimbrites.

ACKNOWLEDGMENTS

This report is based on field data that was collected in the summer of 1991, during which Ron Ryan capably carried out the duties of field assistant and animal control officer. Dr. Abdel Rahman contributed with many discussions in the field. The logistical support from Goose Bay by Wayne Tuttle and Richard White was much appreciated. The manuscript was improved by the comments of C.F. O'Driscoll.

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VOLCANIC METALLOGENIC DATABASE

C.M. Saunders
Mineral Deposits Section

ABSTRACT

Work has begun on a computer database that will hold all available, good-quality, public-domain geochemical data on volcanic rocks for Newfoundland and Labrador. The structure of the database has been completed and data inputting/importing has begun. The data will be held in two main tables; the FIELDDAT table will store descriptive data for each sample such as rock type, location and unit; the GEOCHEM table will hold the analytical results, and related information such as the method of analysis, year of analysis and analytical laboratory. Ancillary tables will hold additional information such as explanations of abbreviations. The tables can be combined so that data from both the main tables can be viewed together. A user will be able to select data by sample number, rock type, unit, NTS, geologist, element or oxide, method of analysis, priority, year of analysis or analytical laboratory or any combination of these factors. Forms for manual input of data and reports for output of data to file or printer have been designed.

INTRODUCTION

During the past year, work began on a geochemical database for volcanic rocks in Newfoundland and Labrador. The structure of the database has recently been completed and the input of data has begun. This paper describes, in general terms, the database structure and methods of inputting/importing and outputting/exporting data.

PROJECT SCOPE AND OBJECTIVES

The aim of the project is compilation of all available good-quality, public-domain geochemical data on volcanic rocks for both insular Newfoundland and Labrador. This will include provincial- and federal-survey data, and data from theses. Analyses from assessment reports may be included if they are found to be of good quality and contain sufficient controls.

Where an element has been analyzed by more than one method, results for each method will be included and data will be ranked according to reliability of the particular method for each element.

The database will provide a repository for a broad spectrum of geochemical data and is particulary aimed at users for whom the geochemistry of volcanic rocks is a useful mineral-exploration tool, i.e., those engaged in the search for base-metal, massive-sulphide deposits.

DATABASE STRUCTURE

The database software is R-base 3.1, a relational database system designed to run on an MS-DOS microcomputer. The data are contained in a series of tables, which can be linked in a variety of ways. Two main tables (Table 1) hold the data—one contains descriptive data for the samples, such as location, rock-type or owner, and the other contains the analytical data. Ancillary tables (e.g., Table 2) contain helpful information, such as preferred listing order of elements and explanations of abbreviations. The tables can be linked so that the user can see data from more than one table at the same time.

Data can be imported from an ASCII file or spreadsheet or can be input manually via a form or directly into the table. It can be exported directly to a file or can be formatted in a report for export to a file or printer.

MAIN TABLES

Table 1 (FIELDDAT) contains descriptive data in the following columns:

COLUMN	TYPE	WIDTH	
Labnum	*integer*	7	
Fieldnum	*text*	8	
Unit	*text*	*	group or complex
Subunit	*text*	*	formation or member
NTS	*text*	5	
UTMeast	*integer*	6	
UTMnorth	*integer*	7	
Zone	*integer*	2	
Rocktype	*text*	4	four-letter code
Owner	*text*	*	project geologist
Age	*text*	*	
Altered	*integer*	1	code for alteration level
Restrict	*text*	1	Y or N to restrict access
Info	*note*	*	comments—e.g., DDH numbers

* indicates that the column width will be expanded as necessary to accommodate text as full names are used rather than codes.

This project is funded by the Canada–Newfoundland Cooperation Agreement on Mineral Development (1990-1994); project carried by Geological Survey Branch, Department of Mines and Energy, Government of Newfoundland and Labrador.

Table 1. Main tables—FIELDDAT contains descriptive data for each sample and GEOCHEM contains the analytial data. Some typical data are shown. Refer to Table 2 for explanations of abbreviations and to text for discussion of columns; -0- indicates a null value.

FIELDDAT

Labnum	Fieldnum	Unit	Subunit	NTS	Zone	Utmeast	Utmnorth	Rocktype	Owner	Age	Restrict	Altered	Info
1543292	BK87014	Lushs Bight Group	Pillow lava member	2E12	21	576220	5494540	BSLT	B.Kean	Ordovician	n	3	Little Bay glory hole
1543006	DE85045	Victoria Lake Group	-0-	12A16	21	576950	5407000	BSLT	D.Evans	Ordovician	n	1	-0-
1543002	DE85007							BSLT	D.Evans	Ordovician	n	1	-0-
etc.													

GEOCHEM

Labnum	Fieldnum	Element	Method	Result	Priority	Labcode	Year	Allmeth
1543292	BK87014	SiO_2	AAS	48.25	1	NDM	1987	AAS
1543292	BK87014	TiO_2	AAS	0.70	1	NDM	1987	AAS
1543292	BK87014	La	ICM	0.93	1	MUN	1989	ICM, ICS
1543292	BK87014	La	ICS	1	2	NDM	1988	ICM, ICS
etc.								

For ease of use, columns such as *Unit*, *Subunit* and *Age* will contain full names rather than abbreviations or codes. The *Rocktype* column contains a four-letter code in order to ensure compatibility with other in-house databases.

Table 1 (GEOCHEM) contains the analytical data in the following columns:

COLUMN	TYPE	WIDTH	
Labnum	*integer*	7	
Fieldnum	*text*	8	
Element	*text*	5	
Method	*text*	3	code for analytical method
Result	*double*	*	variable no. of decimal places
Priority	*integer*	1	rank of analytical method
Labcode	*text*	3	code for analytical laboratory
Year	*integer*	4	year of analysis
Allmeth	*text*	*	list of methods for element

All analyses in Table 1 (GEOCHEM) are contained in one column—*Result*, and the element or oxide is identified by the column *Element*. The column *Priority* provides the user with a quick method of selecting the best data for each sample. This is achieved by assigning a value of 1 to the best method of analysis of an element for a particular sample. For example, Zr by X-ray fluoresence (XRF) might be ranked 1 and Zr by inductively-coupled plasma spectrometry (ICS) ranked 2 for a particular sample; another sample might have been analyzed for Zr by ICS (ranked 1) and by atomic absorption spectrometry (AAS) (ranked 2). Niobium by ICS might be ranked 1 and Nb by AAS, ranked 2. The column *Allmeth* lets the user see at a glance whether an element has been analyzed by more than one method. The user could, therefore, select all priority 1 data or select all data resulting from a particular method.

Because the elements and oxides are listed as values in a column rather than as column headers, each sample number in the GEOCHEM table is repeated as many times as there are individual analyses for that sample (Table 1). In the FIELDDAT table, each sample number exists only once. The FIELDDAT:GEOCHEM tables, therefore, have a one-to-many relationship.

ANCILLARY TABLES

Helpful information is contained in several small subordinate tables (e.g., Table 2). The most important table is called CLASSIFY. This table holds information about elements and oxides and provides a means of controlling the selection and display of geochemical data. It contains the following columns:

Table 2. Ancillary tables that contain helpful information. Abbreviations in *Subcateg* and *Category* columns: TE−transition element, MET−metallic element, LFS−low field strength element, OXI−oxide, MAJ−major element, TRA−trace element. Other abbreviations as shown

CLASSIFY

Element	Atomic#	Weight	Subcateg	Sequence	Printout	Category
Si	14	28.086	-0-	-0-	n	MAJ
Cr	24	51.996	TE	18	y	TRA
Zn	30	65.37	MET	25	y	TRA
Rb	37	85.47	LFS	47	y	TRA
SiO_2	-0-	60.07	OXI	1	y	OXI
etc.						

ROCK

Rocktype	Rockname
ANDS	andesite
BSLT	basalt
DACT	dacite
DIAB	diabase
FLST	felsite
etc.	

ANALYST

Labcode	Labname
BCQ	Becquerel
CHX	Chemex
NDM	Nfld. Dept Mines & Energy
GSC	Geol. Survey of Canada
MUN	Memorial Univ. of Nfld.
etc.	

METHOD

Method	Analmeth
AAS	Atomic Absorption Spectrometry
ICM	Inductively Coupled Plasma-Mass Spectrometry
ICS	Inductively Coupled Plasma Emmission Spectrometry
INA	Instrumental Neutron Activation Analysis
XRF	X-ray Fluorescence
etc.	

COLUMN	TYPE	WIDTH	
Element	*text*	5	
Atomic#	*text*	3	
Weight	*double*	*	atomic or molecular weight
Subcateg	*text*	3	HFS or REE etc.
Sequence	*integer*	2	listing order (SiO_2 = 1)
Printout	*text*	1	Y or N to select element
Category	*text*	3	OXI, MAJ or TRA

Column *Category* lets a user distinguish between major and trace elements; *Subcateg* further subdivides the trace elements, e.g., high-field strength (HFS) or rare-earth elements (REE). Column *Printout* can be used to select or reject each element for output to a file or printer. Column *Sequence* controls the order in which elements are listed in certain reports and applications.

Tables METHOD, ROCK, and ANALYST provide explanations of abbreviations used in the two main tables and ensure consistent use of these codes. Table ROCK for example would contain only 2 columns:

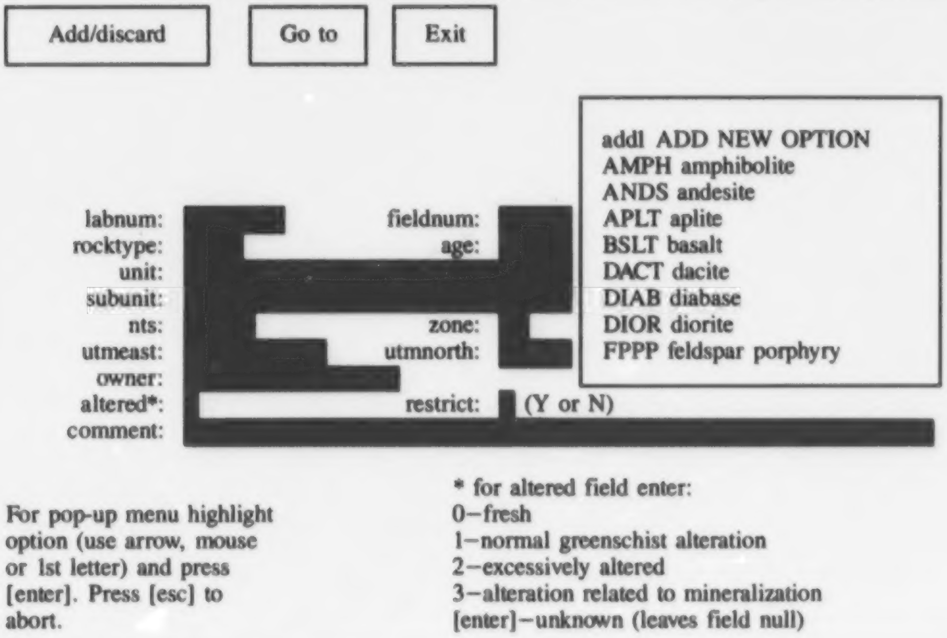
COLUMN	TYPE	WIDTH	
Rocktype	*text*	4	code e.g. BSLT
Rockname	*text*	*	full name e.g. basalt

Tables METHOD and ANALYST similarly contain 2 columns of abbreviations and full names (Table 2).

SELECTING DATA

The two main tables, along with the CLASSIFY table, can be combined into a view that lets the user access all data. A user will be able to select data by sample number, rock type, unit, NTS, element or oxide, method of analysis, priority, year of analysis or analytical laboratory or any combination of these factors. The elements/oxides can be selected on an individual basis or by using the *Category* or *Subcateg* columns of the CLASSIFY table. A user can quickly select all data for a particular sample or can select only priority 1 data or only data that have been analyzed by a particular method. The results of the selection can then be saved into another table or output to another file or to the printer.

Table 3. Form FLDFORM, which is used to input data into the FIELDDAT table. Each field represents a column in the table. (The comment field loads the *Info* column). Shown at right is a pop-up menu that provides the user with the correct abbreviation for each rock type. The selected four-letter code is automatically loaded into the *Rocktype* field. The user can add a new code if needed



For pop-up menu highlight option (use arrow, mouse or 1st letter) and press [enter]. Press [esc] to abort.

* for altered field enter:
0–fresh
1–normal greenschist alteration
2–excessively altered
3–alteration related to mineralization
[enter]–unknown (leaves field null)

IMPORTING DATA

Data can be imported into the database from an ASCII file or from a spreadsheet. Because the structure of the GEOCHEM table differs significantly from standard geological format, data for this table must first be transposed into the correct format. This is most easily accomplished via a spreadsheet. Most Department of Mines and Energy data are already available in SPSS format, which can easily be imported into Lotus 123. Data can then be cleaned up and put into the proper format using a macro. Information such as method of analysis or year of analysis can be entered at this stage or can be entered later into R-Base. The best approach to this depends, to some extent, on the characteristics of each dataset. Some department data exist as separate files divided on the basis of analytical laboratory or year of analysis. Other datasets contain all data for an area regardless of how, when or where the samples were analyzed. The user may have to spend considerable time trying to sort out such details or trying to track down missing data such as UTM coordinates. The cooperation of the project geologist is of great importance at this stage.

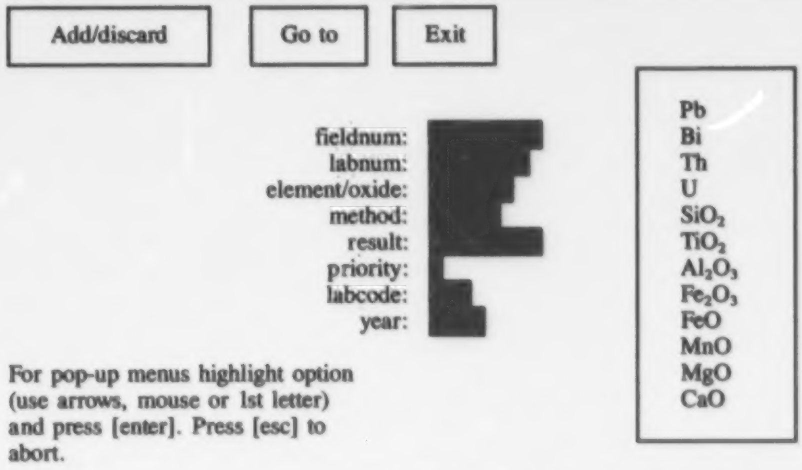
Data can also be exported from R-base into other packages or into ASCII format, however, it is generally necessary to first select the desired information and put it into standard geological format using a report (see Reports section further on).

FORMS

Standardized forms for manual input of data have been designed and will be used where data are not available in electronic format. A form called FLDFORM loads data into the FIELDDAT table. The *Rocktype* field has an automatic pop-up menu that provides the user with the proper abbreviation for each rock type (Table 3). The user can add a new option to this pop-up menu if the desired rock type is not already listed.

Another form called CHEMFORM (Table 4) loads analytical data into the GEOCHEM table. To speed up the input of data, the form can be set up so that certain fields contain default values, e.g., if all analyses to be typed in were acquired in 1988, the Year field could be set to this value. A field can also be set to duplicate the value in the previous row so, for example, the sample numbers would only have to be typed in once for each sample. A pop-up menu in the CHEMFORM form (Table 3) lists elements and oxides to speed their input into the *Element* field in the GEOCHEM table. Only the elements of interest are displayed. This is accomplished by use of a form called PRINFORM that loads the *Printout* column in the CLASSIFY table with either 'y' to include an element or 'n' to exclude it. This can also be used to select elements for inclusion in views and reports.

Table 4. Form CHEMFORM, which is used to input data into the GEOCHEM table. The pop-up menu at right speeds up the entry of element/oxide names



For pop-up menus highlight option (use arrows, mouse or 1st letter) and press [enter]. Press [esc] to abort.

REPORTS

Two basic types of reports have been designed. The first, called LISTREP, produces a one-sample per-page report (Table 5) that contains all selected geochemical data from the GEOCHEM table, as well as descriptive data from the FIELDDAT table, such as rock type, unit and location. The driving view for this report is called PRINVIEW and combines the tables GEOCHEM, FIELDDAT, and CLASSIFY. Data can be restricted by any column, e.g., *Unit, NTS, Rocktype, Priority,* etc., by building this view to contain only the desired data or by using quick select from the report printing menu. Elements can be selected on an individual basis by using the form PRINFORM or on a group basis by using the columns *Category* or *Subcateg*. The data would ideally be sorted from the report printing menu, e.g., by *Unit, Subunit, Rocktype, Labnum* and *Sequence*. Sorting by *Sequence* is necessary if the user wishes to have the components listed in standard order (SiO_2, TiO_2, Al_2O_3, etc.). Sorting by *Element* would not work as this would result in an alphabetical listing (Ag, Al_2O_3, As, etc.). This report presents both hydrous and anhydrous data and lists the method and year of analysis and the analytical laboratory. The column *Allmeth* is also listed to let the user see if any additional data exist for a particular element.

The second type of report called ARRAYREP produces a listing in standard geological format (Table 6) for export to programs such as IGPET. For this report, data must be transformed by use of variables (lookups) to achieve the desired format; as a result, if the number of elements is greater than about 23, more than one report must be used to avoid memory problems. In practice it would be most convenient to put all major elements in one report and all trace elements in another report. If the data are to be printed, the number of elements per report will be limited by the width of the paper. If the data are to be exported to IGPET, the user must first decide what method of analysis (normally priority 1) to select for each element as IGPET can only hold one analysis per sample for each element. The driving table for this report is the FIELDDAT table; this table is required because it contains only one occurrence of each sample number in the Labnum column. Data can be limited by, or sorted by, columns within this table at the report printing menu. However, in order to limit data by method, laboratory, priority etc., a temporary table called PRINTTAB must be built. This table will hold a subset of the GEOCHEM table and will significantly reduce processing time. The lookup variables that hold the data are expressions such as:

SiO_2=result in Printtab where labnum=labnum and element=SiO_2

Each column in the resulting report is the product of such an expression.

DISCUSSION

This database has been designed to be flexible and easy to use. The most difficult problem has been how to provide the user with all pertinent information that is necessary to evaluate the quality and comparability of analyses. This includes such parameters as method of analysis, year of analysis and analytical laboratory. An additional complicating factor is the analysis of some samples for a particular element by more than one method. The need to accommodate these factors was the reason for the unconventional format of the GEOCHEM table. This format has two main disadvantages:

- importing/exporting to and from other packages is more complex because existing electronic data are generally in standard geological format or are required to be for input to packages such as IGPET; and

- the number of rows in the GEOCHEM table is greatly increased; this combined with the duplication

Table 5. Sample output from LISTREP report, which provides a one-sample-per-page printout. Both raw and recalculated (anhydrous) data are shown. In this example, AAS and ICS analyses were selected. The *Allmeth* column lets the user know that several elements have also been analyzed by XRF. See Table 2 for codes

Fieldnum: SS015
Labnum: 1341051
Rocktype: trondhjemite
NTS: 12H10

Unit: Gull Lake intrusive suite
Subunit: Black Duck Ponds trondhjemite
Geologist: W.R. Smyth
UTMeast: 506200 UTMnorth: 5498750

Component	Result	Recalc	Meth	Lab	Year	AllMeth*
SiO_2	74.8	75.41	AAS	NDM	1981	AAS
TiO_2	0.31	0.31	AAS	NDM	1981	AAS
Al_2O_3	12.8	12.90	AAS	NDM	1981	AAS
Fe_2O_3	1.29	1.30	AAS	NDM	1981	AAS
FeO	1.84	1.85	AAS	NDM	1981	AAS
MnO	0.06	0.06	AAS	NDM	1981	AAS
MgO	0.72	0.72	AAS	NDM	1981	AAS
CaO	1.16	1.16	AAS	NDM	1981	AAS
Na_2O	5.4	5.44	AAS	NDM	1981	AAS
K_2O	0.76	0.76	AAS	NDM	1981	AAS
P_2O_5	0.05	0.05	AAS	NDM	1981	AAS
LOI	1.41	1.42		NDM	1981	
Total	100.6	101.42		NDM		
Cu	1.	1.00	AAS	NDM	1982	AAS
Pb	1.	1.00	AAS	NDM	1982	AAS
Zn	27.	27.22	AAS	NDM	1982	AAS
Ag	1.	1.00	AAS	NDM	1982	AAS
Rb	12.	12.09	AAS	NDM	1982	AAS, XRF
Ba	55.	55.44	AAS	NDM	1982	AAS, ICS, XRF
Ba	79.	79.64	ICS	NDM	1988	AAS, ICS, XRF
Sr	68.	68.55	AAS	NDM	1982	ICS, XRF, AAS
Sr	80.	80.65	ICS	NDM	1988	ICS, XRF, AAS
Li	24.	24.19	AAS	NDM	1982	AAS
Nb	4.	4.03	ICS	NDM	1988	ICS, XRF
Zr	71.	71.57	ICS	NDM	1988	XRF, ICS
Y	13.	13.10	ICS	NDM	1988	ICS, XRF
Th	0.	0.00	ICS	NDM	1988	ICS, XRF

* Analytical methods in *Allmeth* column listed in order of priority.

of sample numbers in the key *Labnum* column slows down searches and sorts.

However, there are many advantages to the format:

- all pertinent information is included;
- the number of columns is significantly reduced; an alternative structure (separate columns for each method of analysis of each element) would result in an unwieldy number of columns;
- the amount of wasted (null) space is greatly reduced; the alternative structure described above would be inefficient in that if only 50 samples out of several thousand had been analyzed for Co by atomic absorption most of the Co-AAS column would contain nulls;
- data may be selected or sorted by any number of parameters such as NTS, unit, rock type, method of analysis, etc;
- it is easy to determine, without any prior knowledge, what type of data exist for any particular sample, or for any particular area or rock unit; and
- this structure is easily adapted for use with other types of data, e.g. soil geochemistry.

Table 6. Sample output from ARRAYREP report, which lists data in standard geological format. Priority 1 data have been selected. Generally, major elements would be included in one report and trace elements in another report as there is a limit of about 23 elements due to memory limitations. The dashed line indicates a null value. The user can choose the method of displaying nulls

'Sample'	'SiO_2'	'TiO_2'	'Al_2O_3'	'Fe_2O_3'	'LOI'	'Cr'	'Cu'	'Zn'	'Ba'	'Rb'
1543001	61.20	0.95	14.87	1.25	3.84	9	702	128	28	9
1543002	52.90	0.59	14.77	0.82	7.71	199	59	130	189	7
1543004	83.45	0.13	8.40	----	0.51	4	27	17	338	22
1543009	47.40	2.06	16.78	1.34	3.36	134	68	76	51	7
1543011	46.90	2.00	16.17	1.58	2.94	174	68	90	58	8
1543012	45.60	2.06	15.53	2.11	3.47	176	75	90	42	11
1543013	46.70	2.25	16.88	2.38	2.65	53	62	90	54	15
1543014	47.40	1.73	15.95	1.33	2.87	215	71	76	48	13

FUTURE CONSIDERATIONS

Although the structure is essentially complete, there are other problems yet to be solved such as the method of handling controls (standards and duplicates) and the establishment of a methodology for evaluating, assessing and reporting relative data quality. This will be more easily accomplished after a considerable amount of data has been loaded into the database. Other parameters may be further modified or refined. For example, the method of analysis as currently coded gives no indication of the type of sample dissolution that was used. This could be indicated by further refinement of the codes used or by addition of another column to hold such information. The GEOCHEM table may eventually have to be subdivided into smaller tables based on geographic regions, such as 1:250 000 or 1:50 000 map sheets because of the large number of rows. Longer term considerations include design of applications to make querying or adding data easier for those not familiar with R-base. These problems will require consultation with other members of the Department and other industry, government and university personnel—questions, comments and helpful hints are welcomed.

ACKNOWLEDGMENTS

Jim Patterson suggested the structure of the GEOCHEM table and has been of assistance with general trouble-shooting; he was instrumental in the solving of initial memory problems. Scott Swinden provided guidance concerning the scope and objectives of the project. Jim Butler has helped ensure compatibility with other in-house databases. This project is funded by the Canada–Newfoundland Cooperation Agreement on Mineral Development, 1990-1994.

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VOLCANIC STRATIGRAPHY NORTHWEST OF NEW BAY POND, CENTRAL NEWFOUNDLAND, AND THE STRIKE–EXTENT OF THE POINT LEAMINGTON MASSIVE SULPHIDE HORIZON

H.S. Swinden and G.A. Jenner[1]
Mineral Deposits Section

ABSTRACT

Geological mapping in the area west and north of New Bay Pond, central Newfoundland, suggests that submarine volcanic and epiclastic rocks informally assigned to the New Bay Pond sequence may be correlated with the Wild Bight Group immediately to the north. Whole-rock geochemical data support this correlation and allow the New Bay Pond sequence lavas to be interpreted in the context of the chemostratigraphy of the Wild Bight Group. Pillow lavas that form the northeastern unit of the New Bay Pond sequence consist mainly of incompatible element-depleted rocks with unusual REE-patterns. These rocks are similar to those found in the central part of the Wild Bight Group stratigraphic section where they are invariably associated with volcanogenic sulphide deposits. In contrast, the southwestern volcanic unit in the New Bay Pond sequence consists of light REE-enriched tholeiitic and mildly alkalic rocks. They are similar to rocks at the top of the Wild Bight Group, which are not known to be associated with mineralization.

Interpretation of geochemical and geophysical (electromagnetic and vertical gradient magnetometer) data suggest that the highly incompatible element-depleted rocks in the New Bay Pond sequence form part of a geological unit that can be traced continuously northward to the area of the Point Leamington massive sulphide deposit. The data indicate that this unit and its contained massive sulphide deposit are separated from volcanic and sedimentary rocks to the east by a thrust fault that predates regional folding. Numerous northeast-striking faults in the area have small displacements and do not significantly disrupt the stratigraphic succession.

INTRODUCTION

Submarine mafic and felsic volcanic rocks and volcaniclastic turbidites between New Bay Pond and Frozen Ocean Lake (Figures 1 and 2) were assigned by Dean (1978) to the Frozen Ocean Group, which he considered to be of probable Silurian age. Kusky (1985) and Swinden (1988a) independently mapped this area and reinterpreted these rocks as pre-Middle Ordovician correlatives of the Wild Bight Group, a thick sequence of dominantly volcanic-derived turbidites interbedded with mafic and minor felsic volcanic rocks immediately to the north and northeast (Figure 2). Swinden (1988a) suggested that the stratigraphic name 'Frozen Ocean Group' be abandoned, and informally coined the term 'New Bay Pond sequence' for the marine volcanic and sedimentary rocks northwest of New Bay Pond.

Although the overall lithological similarity between the New Bay Pond sequence and the Wild Bight Group suggest that the two might be correlative, this interpretation is equivocal. There are many other marine mafic volcanic–volcaniclastic sequences in central Newfoundland, which are lithologically similar to, and thereby also potentially correlative with, the New Bay Pond sequence, although none are so closely associated. One way of testing the correlation is through study of the whole-rock geochemistry of the mafic volcanic rocks. Such an approach is feasible because the chemostratigraphy of the Wild Bight Group has been particularly well-documented and it has been shown that geochemically distinct mafic volcanic rocks are characteristic of particular intervals within the stratigraphic sequence (Swinden, 1987; Swinden *et al.*, 1990). Therefore, it is likely that if the mafic volcanic rocks in the New Bay Pond sequence do correlate with parts of the Wild Bight Group, their chemostratigraphic association can be recognized. Of particular interest, in the context of this study, is whether the stratigraphic interval that hosts the large Point Leamington massive sulphide deposit immediately to the northwest of the New Bay Pond area could be traced to, or recognized in, the New Bay Pond sequence, thereby enhancing the exploration potential of this area.

Ten samples of mafic volcanic rocks were collected from the New Bay Pond sequence and analyzed for a

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[1] Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland

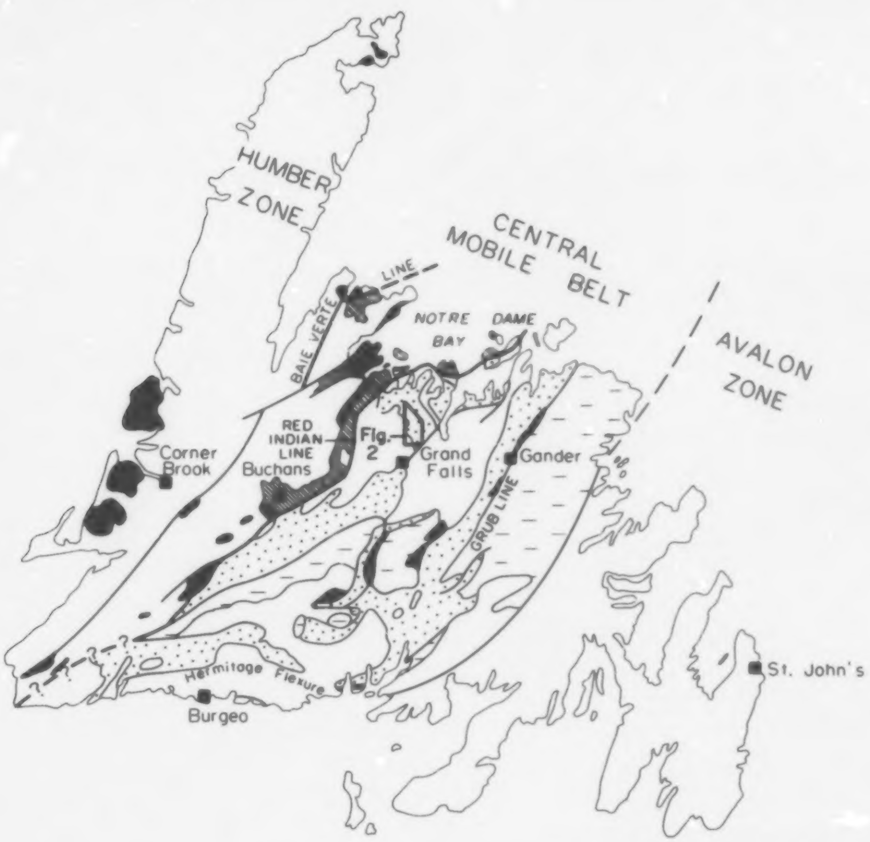


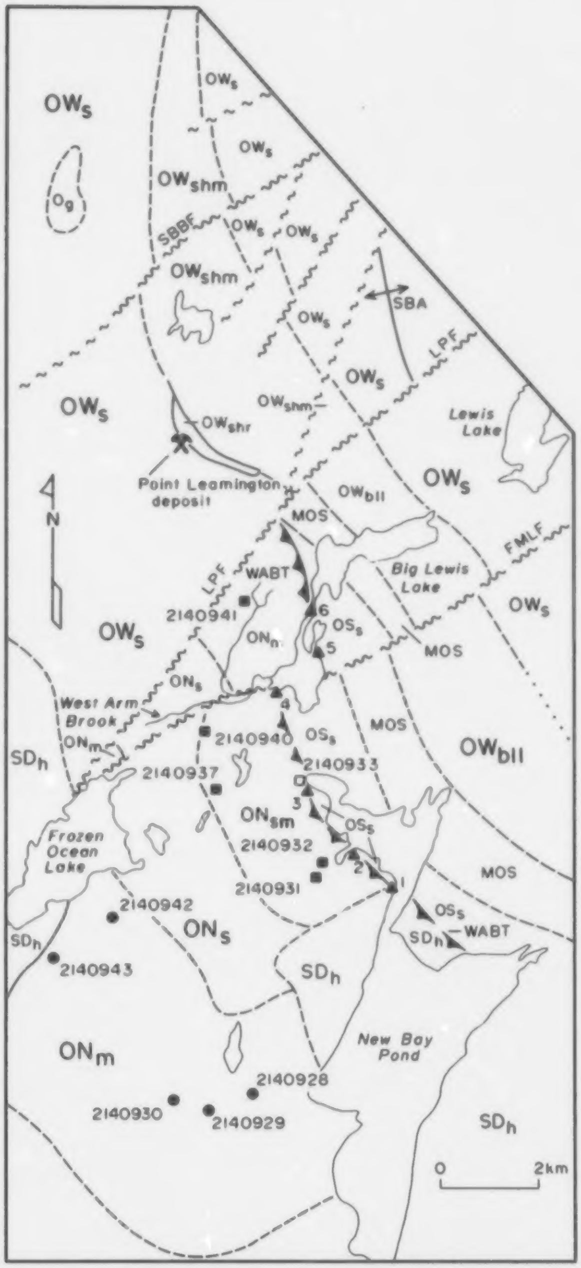
Figure 1. *Location map showing vestiges of Iapetus in central Newfoundland (after Swinden* et al., *1988). Ophiolites are in black. Hatched and stippled areas are volcanic and epiclastic rocks in the Notre Dame and Exploits subzones (Williams* et al., *1988), respectively. Solid box outlines area of Figure 2.*

comprehensive suite of major and trace elements (Figure 2; Table 1). These data are herein used to examine the proposed lithostratigraphic correlations between the New Bay Pond sequence and the Wild Bight Group and, coupled with results from gradiometer (Geological Survey of Canada data) and industrial airborne electromagnetic surveys, to revise the geological map of the area between the Point Leamington deposit and New Bay Pond. New interpretations resulting from these data suggest that the productive Point Leamington stratigraphic interval can be traced intact from the deposit southward to the area of New Bay Pond and that a structural model involving early, south-directed thrusting can explain previously unresolved stratigraphic problems in the area of the massive sulphide deposit.

GENERAL GEOLOGY

The early Paleozoic geological history of the central part of Notre Dame Bay is recorded by a structurally disrupted sequence of Cambrian to Ordovician volcanic and volcaniclastic rocks, which record events in a number of island arcs and back-arc basins. The Wild Bight and adjacent Exploits groups preserve the youngest arc and back-arc sequences in Newfoundland (Swinden *et al.*, 1989a; 1990) and are regionally overlain by a Middle Ordovician chert and carbonaceous argillite unit (the 'Caradocian Interval' of Dean, 1978), which passes gradationally and conformably into Upper Ordovician and Lower Silurian volcanic-derived turbidites (the Sansom and Point Leamington greywackes and the Goldson Conglomerate).

The geology of the Wild Bight Group has been most recently described by Dean (1978) and Swinden (1987). It is disposed about a major, north-trending fold, the Seal Bay anticline. Rocks in the core of the Seal Bay anticline represent the stratigraphically lowest units in the Wild Bight Group and the stratigraphic succession progresses more or less continuously upward across the limbs of this fold (although with local reversals resulting from folding and faulting) until passing into the overlying Shoal Arm Formation ('Caradocian interval'). Rocks in the Wild Bight Group comprise about 25 percent volcanic rocks, of which aqpproximately 90 percent are mafic pillow lavas and lesser massive basalts. There are four volcanogenic sulphide deposits or occurrences in the central part of the Wild Bight Group, the Point Leamington and Lockport deposits and the Indian Cove and Long Pond


OWs
OWs
OWs
Og
OWshm
SBBF
OWs
OWs
OWshm
SBA
OWs
LPF
OWs
OWs
OWshm
OWshr
Lewis Lake
Point Leamington deposit
OWbll
OWs
N
MOS
FMLF
WABT
Big Lewis Lake
LPF
2140941
6
OSs
OWs
OWs
ONn
5
MOS
ONs
West Arm Brook
4
MOS
OWbll
ONm
2140940
OSs
SDh
2140937
2140933
3
ONsm
OSs
Frozen Ocean Lake
2140932
MOS
2
2140931
1
2140942
OSs
ONs
SDh
WABT
SDh
2140943
SDh
ONm
New Bay Pond
2140928
2140930
2140929
SDh
0
2km

LEGEND

SILURIAN OR DEVONIAN

SD_h	Hodges Hill Granite–*dominantly granite, lesser mafic intrusive rocks*

ORDOVICIAN–SILURIAN

OS_s	Sansom Greywacke–*dominantly volcanically-derived turbiditic greywacke, lesser conglomerate*

MIDDLE ORDOVICIAN

MOS	Shoal Arm Formation–*carbonaceous shale, chert*

ORDOVICIAN

Wild Bight Group

OW_s	*Sedimentary rocks, dominantly volcanically-derived sandstone turbidites, lesser argillite, conglomerate*
OW_{shm}	*Side Harbour volcanic unit pillow lavas*
OW_{shr}	*Side Harbour volcanic unit rhyolite*
OW_{bll}	Big Lewis Lake basalt–*pillow lava, hyaloclastic mafic breccia*

New Bay Pond sequence

NB_m	*Pillow lava, hyaloclastic mafic breccia*
NB_s	*Sedimentary rocks, dominantly volcanically-derived turbidites, lesser argillite, conglomerate*

SYMBOLS

∿∿	*Strike slip fault*
▲▲	*Thrust fault; teeth point to hanging wall*
/ / ⁄	*Geological contact; defined, approximate, assumed*
‡	*Anticline*
⚒	*Volcanogenic massive sulphide deposit*

Figure 2. *General geology in the area of New Bay Pond and the Point Leamington deposits sensu Dean (1978) with revisions (after Kusky, 1985), Swinden (1987) and Swinden (1988a). SBBF–Seal Bay Brook Fault; LPF–Long Pond Fault; FMLF–Four Mile Lake Fault; WABT–West Arm Brook Thrust; SBA–Seal Bay Anticline. Solid triangles indicate locations of descriptions of the West Arm Brook Thrust (Kusky, 1985). Analyzed sample locations with sample numbers are keyed to geochemical type (see text): solid squares–Type I; open squares–Type II; solid dots–Type III.*

deposits (Swinden, 1988b). The largest of these, the Point Leamington deposit, contains about 13.8 million tonnes of massive sulphide grading 0.48 percent Cu, 1.92 percent Zn, 18.1 g/t Ag and 0.9 g/t Au (Walker and Collins, 1988). The Lockport deposit is a small, massive-sulphide lens having an extensive footwall alteration zone, whereas the Indian Cove and Long Pond occurrences are stockworks.

In the New Bay Pond area, the top of the Wild Bight Group consists of a thick unit of pillow lava and pillow breccia

Table 1. Geochemical analyses of New Bay Pond Sequence mafic volcanic rock

Sample	2140928	2140929	2140930	2140931	2140932	2140933	2140940	2140941	9140942	9140943
SiO_2	51.65	49.05	48.90	49.30	52.40	51.25	55.40	62.30	48.30	50.25
TiO_2	1.72	1.97	1.50	0.16	0.32	0.68	0.48	0.33	1.34	1.81
Al_2O_3	17.29	15.15	14.30	12.92	17.68	14.86	14.35	12.62	12.04	13.77
Fe_2O_3	5.06	4.96	6.13	0.78	1.63	4.52	5.01	1.67	2.12	2.74
FeO	3.25	6.46	4.62	7.96	8.04	9.03	7.00	6.39	8.19	7.82
MnO	0.21	0.18	0.19	0.15	0.11	0.19	0.13	0.13	0.20	0.16
MgO	5.07	7.02	6.45	8.68	6.99	4.89	3.69	4.06	12.47	9.78
CaO	9.82	9.67	14.01	9.62	3.78	8.77	10.25	5.21	10.66	8.47
Na_2O	4.11	3.41	2.65	0.85	5.54	2.81	0.47	4.02	2.26	3.80
K_2O	0.73	1.09	0.13	1.01	0.07	0.20	0.07	0.11	0.17	0.35
P_2O_5	0.36	0.25	0.19	0.01	0.01	0.05	ND	ND	0.18	0.21
Total	100.66	100.52	100.60	99.53	100.20	99.78	99.65	100.69	100.18	100.13
LOI	1.39	1.31	1.53	8.09	3.63	2.50	2.81	3.12	2.25	0.97
Mg #	54	53	53	64	57	40	36	48	69	63
Cr	71	114	366	729	101	7	7	25	805	463
Ni	35	56	113	118	34	8	14	14	374	196
Sc	NA	29	25	NA	51	41	58	NA	30	31
V	205	268	260	252	306	424	425	271	213	230
Cu	9	32	55	127	102	58	106	154	80	89
Zn	81	97	80	75	77	97	86	69	79	91
Rb	15	23	ND	14	6	7	ND	ND	ND	7
Sr	550	315	215	108	73	62	122	28	313	220
Ta	NA	0.74	0.56	NA	ND	0.03	ND	NA	0.53	0.73
Nb	NA	13.9	10.1	NA	0.2	0.4	0.2	NA	9.2	13.1
Hf	NA	3.10	2.48	NA	0.13	0.65	0.21	NA	2.58	–
Zr	172	117	81	NA	4	18	5	1	98	125
Y	NA	20	17	NA	3	15	4	NA	17	20
Th	NA	2.11	1.11	NA	0.04	0.25	0.09	NA	1.50	1.57
La	NA	12.60	9.86	NA	0.28	1.51	0.51	NA	9.71	11.50
Ce	NA	28.60	21.80	NA	0.54	3.56	0.79	NA	21.70	26.70
Pr	NA	3.90	2.90	NA	0.11	0.64	0.20	NA	3.06	3.93
Nd	NA	17.40	12.60	NA	0.65	3.17	0.88	NA	13.60	17.00
Sm	NA	4.61	3.34	NA	0.28	1.21	0.32	NA	3.55	4.37
Eu	NA	1.52	1.20	NA	0.11	0.52	0.18	NA	1.07	1.57
Gd	NA	4.74	3.90	NA	0.39	1.75	0.50	NA	3.96	4.82
Tb	NA	0.72	0.56	NA	0.08	0.33	0.09	NA	0.57	0.69
Dy	NA	4.20	3.34	NA	0.61	2.49	0.74	NA	3.54	4.02
Ho	NA	0.82	0.65	NA	0.13	0.56	0.17	NA	0.68	0.74
Er	NA	2.17	1.81	NA	0.46	1.79	0.50	NA	1.84	2.03
Tm	NA	0.30	0.24	NA	0.06	0.26	0.08	NA	0.25	0.31
Yb	NA	1.82	1.52	NA	0.40	1.78	0.50	NA	0.63	1.64
Lu	NA	0.25	0.21	NA	0.08	0.28	0.08	NA	0.24	0.25

NA – not analyzed; ND – not detected; major elements in wt. %; trace elements in ppm

informally termed the Big Lewis Lake basalt by Swinden (1987). The Big Lewis Lake basalt is overlain, apparently conformably although the contact has not been observed, by the fossiliferous Caradocian Shoal Harbour Formation. Although there are no direct data as to the age of the Wild Bight Group, its relationship with the Shoal Arm Formation and correlations with the fossiliferous Exploits Group to the east suggest that the top is probably Llanvirn–Llandeilo in age. The Shoal Arm Formation is stratigraphically overlain by Middle Ordovician to Silurian volcanic-derived turbidites of the Sansom Greywacke.

In the area between New Bay Pond and Big Lewis Lake (Figure 2), the Sansom Greywacke is structurally overlain by pillow lavas of the New Bay Pond sequence (Kusky, 1985).

The geology of the New Bay Pond sequence (Figure 2) has been previously described in some detail by Kusky (1985) and Swinden (1988a). The sequence consists of a northwest-striking succession of marine, dominantly mafic, volcanic and volcanic-derived sedimentary rocks. Three lithostratigraphic units are recognized (Figure 2):

1) a basal (southwestern) mafic volcanic unit consisting predominantly of massive basalt and mafic volcanic breccia;

2) an intermediate epiclastic turbidite unit consisting predominantly of greywacke and pebbly conglomerate with lesser argillite and coarse, polylithic conglomerate; and

3) an upper (northeastern) mafic volcanic unit consisting dominantly of pillow lava and recognizable pillow breccia containing minor silicic volcanic rocks. Red, ferruginous chert having interbedded magnetite are present locally.

Facing directions in the epiclastic unit are dominantly to the northeast but contacts between lithostratigraphic units were not observed and it is not known whether the lithological succession is stratigraphic or structural.

Kusky (1985) was the first to recognize that the contact between the New Bay Pond sequence and the Sansom Greywacke to the northeast is a fault, which he informally named the FOG fault. Because this name has not been formally published and does not correspond to accepted practice in structural nomenclature, the fault is herein renamed the West Arm Brook Thrust. Kusky (1985) made detailed structural observations, including measurement of stretching lineations, C−S fabrics, rotated clasts and sheath folds, at 6 localities between New Bay Pond and Big Lewis Lake (Figure 2). From these, he interpreted the kinematic sense of movement on the fault, when rotated to account for later folding, to be consistent with thrusting from north to south. Kusky (1985) suggested that this may have occurred during back thrusting associated with continued convergence of outboard terranes against the Laurentian margin following the Taconian orogenic event.

Dean's (1978) interpretation of stratigraphic and structural relationships northwest of New Bay Pond require a major structural discontinuity in the area north of Big Lewis Lake. This is because the sequence exposed on the shores and south of Big Lewis Lake exhibits a continuous stratigraphic section from the Big Lewis Lake basalt into the Shoal Arm Formation and thence, the Sansom Greywacke, with the Big Lewis Lake basalt representing the stratigraphic top of the Wild Bight Group. Northwest of Big Lewis Lake, however, volcanic rocks that are more or less along strike with the Big Lewis Lake basalt (the Side Harbour volcanic unit of Swinden, 1987 and Swinden *et al.*, 1990) were interpreted to pass stratigraphically upward into a continuing thick section of Wild Bight Group turbidites (Figure 2). In this interpretation, these volcanic rocks must have originated substantially below the Big Lewis Lake basalt, the Side Harbour and Big Lewis Lake basalts could not be stratigraphically correlative, and the fact that they are more or less along strike would be considered to be fortuitous. Such an interpretation necessitates a major structural break between the two volcanic units; Dean (1978) postulated that this break was the northeast-striking Long Pond Fault (Figure 2), requiring a major left-lateral offset on this structure.

GEOCHEMISTRY

THE WILD BIGHT GROUP

Swinden (1987) and Swinden *et al.* (1990) showed through a detailed geochemical and isotopic study that that volcanic rocks in the Wild Bight Group record a transition from island-arc-related volcanism (lower and middle stratigraphic levels) to dominantly non-arc-related volcanism (upper stratigraphic levels). Mafic volcanic rocks with island-arc geochemical signatures (positive Th and negative Nb, Ta anomalies on extended rare-earth plots; see Swinden *et al.*, 1989a, 1990, for discussions of arc and non-arc geochemical signatures) occupy the lower two-thirds of the stratigraphic section, overlain by rocks in the upper third of the section that lack island-arc signatures (Figure 3). Swinden *et al.* (1990) recognized six types of mafic volcanic rocks based on their geochemical characteristics. The lowermost mafic volcanic rocks in the Wild Bight Group consist mainly of light rare-earth-element (LREE)-enriched arc tholeiites and calc-alkalic basalts (Type A-I; this and subsequent references to geochemical type follow the nomenclature of Swinden *et al.*, 1990). They pass upward in the intermediate stratigraphic levels of the group into an interval marked by the eruption of LREE-depleted arc tholeiites (Type A-II), refractory tholeiites of highly distinctive chemical character (Type A-II), and LREE-depleted rhyolites. These pass upward into successions in the upper part of the Wild Bight Group, in which the volcanic rocks are characterized by alkalic basalts (Type N-I), high-TiO_2 tholeiites of oceanic island character (Type N-II) and basalts typical of tholeiites found on mid-ocean ridges (Type N-III).

Swinden *et al.* (1990) proposed a tectonic model for the Wild Bight Group, whereby an early episode of arc volcanism (recorded by arc tholeiites in the lower part of the group) was succeeded by a rifting of the arc (recorded by arc tholeiites, refractory arc tholeiites and rhyolites) in the central part of the group. Following rifting of the arc, a back-arc basin was established, recorded by non-arc rocks in the upper part of the group. The four volcanogenic sulphide deposits in the Wild Bight Group occur in the central part of the stratigraphy and are closely associated with Type A-III refractory tholeiites and rhyolite. Swinden *et al.* (1989a,b) have proposed that this association of massive sulphides with refractory mafic volcanic rocks is characteristic of many of the ancient island-

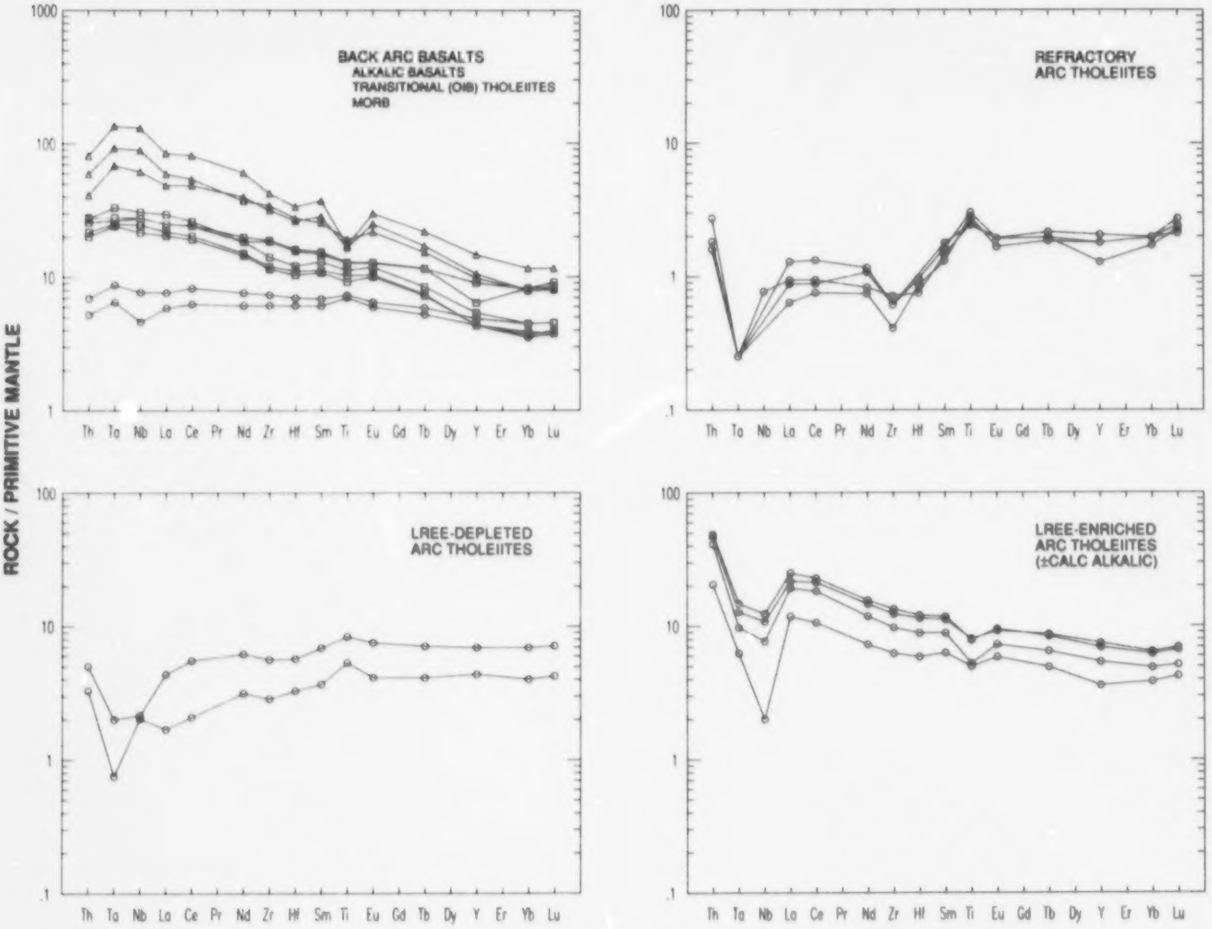


Figure 3. *Chemostratigraphy of the Wild Bight Group (after Swinden* et al., *1990). LREE-enriched arc tholeiites (Type A-I) are found in approximately the lower third of the sequence and are joined by LREE-depleted arc tholeiites (Type A-II) and refractory arc tholeiites (Type A-III) as well as rhyolites (not shown) in the intermediate part of the sequence. Back-arc basalts (triangles–Type N-I; squares–Type N-II; circles–Type N-III) dominate the upper part of the sequence.*

arc environments in central Newfoundland (e.g., Rambler, the Lushs Bight Group, the Betts Cove Ophiolite Complex; Swinden *et al.*, 1989a,b).

Swinden (1987) pointed out that there is a stratigraphic problem in the area of the Point Leamington deposit, if island-arc rocks are supposed to stratigraphically underlie non-arc rocks in the Wild Bight Group. In this area (Figure 2), the Side Harbour volcanic unit appears to be chemically bipartite. The Point Leamington massive sulphide deposit and its associated rhyolite form the western edge of the volcanic unit and are associated with arc-related refractory tholeiites in the immediate footwall. However, pillow lavas less than 1 km to the east and northeast of the deposit, and apparently stratigraphically below by virtue of their relative positions on the west limb of the Seal Bay anticline, are high-TiO_2 non-arc tholeiites and alkalic basalts. Swinden (1987) recognized this apparent anomaly in the stratigraphic relationships and suggested that it might be explained by repetition of stratigraphy along a roughly bedding-parallel fault between the Point Leamington deposit and the volcanic rocks to the east. However, exposure is very poor in the area and there were no geological observations to support or refute this suggestion.

THE NEW BAY POND SEQUENCE

New analyses of mafic volcanic rocks from the New Bay Pond sequence are summarized in Table 1 and illustrated in Figures 4 to 6. Samples were taken from the interior of pillows or the relatively fresh interiors of massive flows. Data for the major elements and the trace elements Cu, Pb, Zn, Co, Ni, V, Ba, Sr and Rb were acquired by inductively coupled plasma–emission spectrometry (ICP-ES) in the Geological Survey Branch Laboratory using a HF, $HClO_4$ and HCl digestion. Analyses for the REE, certain high-field strength elements (HFSE; Ta, Nb, Zr, Hf, Y) and Th were analyzed by inductively coupled plasma–mass spectrometry (ICP-MS) at the Department of Earth Sciences, Memorial University; analytical techniques for ICP-MS with accuracy and precision are reported by Jenner *et al.* (1990).

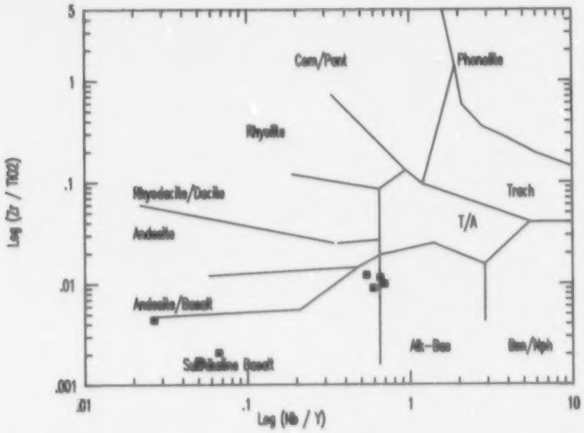


Figure 4. *Discrimination of volcanic magma types for old altered rocks after Winchester and Floyd (1977). New Bay Pond sequence basalts form two groups with affinities to subalkalic basalts and transitional alkalic basalts; respectively.*

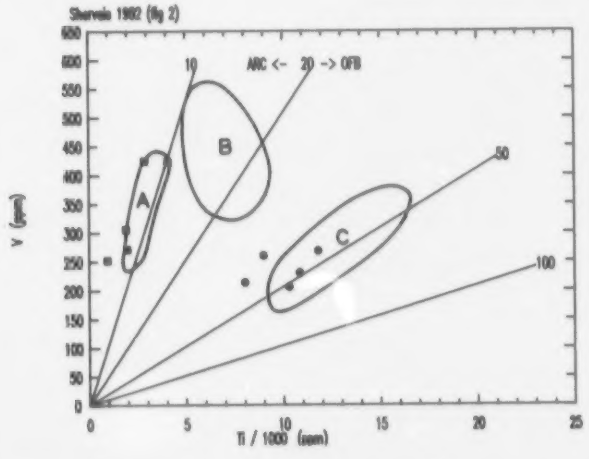


Figure 5. *Ti:V plot for New Bay Pond sequence mafic volcanic rocks after Shervais (1982). Solid squares–Type I; open squares–Type II; solid dots–Type III. Field A–Type A-III refractory tholeiites in the Wild Bight Group; Field B–Type A-II LREE-depleted arc tholeiites in the Wild Bight Group; Field C–Type N-II tholeiites of oceanic island affinity from the Wild Bight Group.*

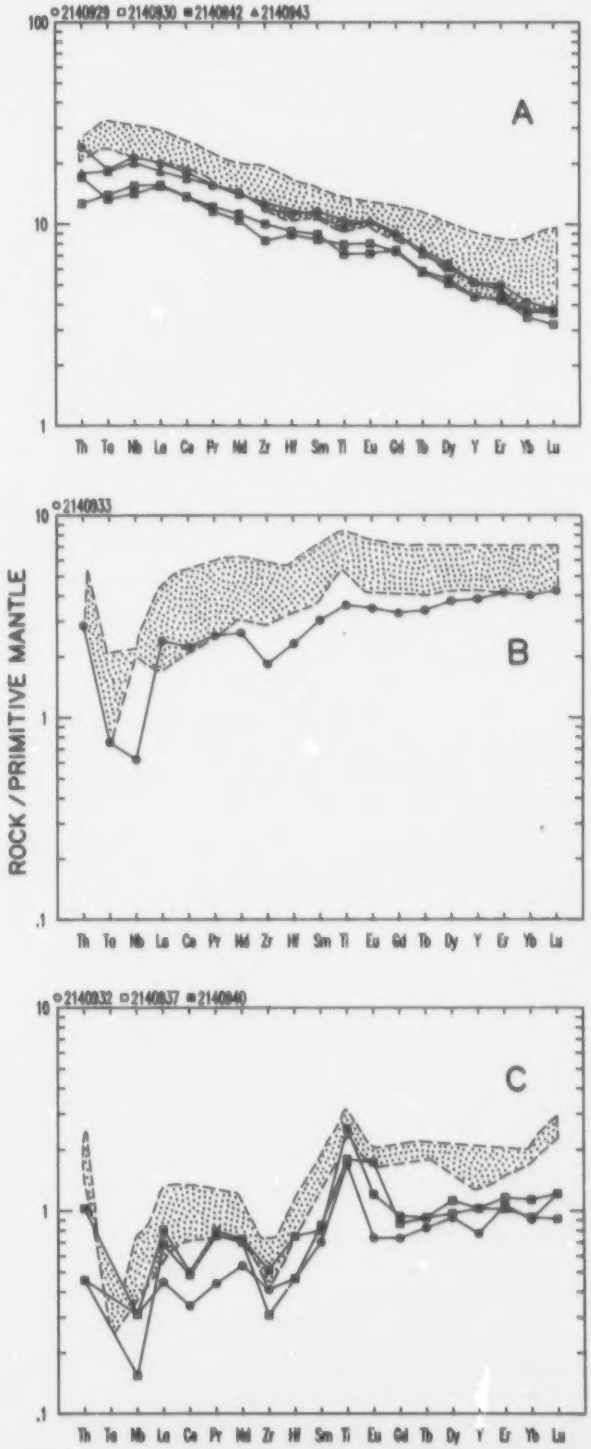


Figure 6. *Extended rare-earth plots for the New Bay Pond mafic volcanic rocks. A–Type I. Stippled field is the Type A-III refractory tholeiites in the Wild Bight Group; B–Type II. Stippled field is Type A-II LREE-depleted arc tholeiites in the Wild Bight Group; C–Type III. Stippled field is Type N-II tholeiites of oceanic island affinity from the Wild Bight Group.*

Mafic volcanic rocks in the New Bay Pond sequence clearly fall into two distinct magma types (Figure 4), a set of subalkalic basalts (Nb/Y <.08) and a set of subalkalic to transitional alkalic basalts (Nb/Y ~ 0.4 to 0.8). Incompatible-element contents and trace-element ratios (illustrated by the Ti:V ratio, Figure 5) support this subdivision and extended REE plots suggest that the subalkalic basalts can be further subdivided into two groups containing lower and higher incompatible-element contents, respectively. The geochemistry of these samples is described below in terms of these three geochemical types.

Type I

Type I mafic volcanic rocks are subalkalic and highly depleted in the incompatible elements (e.g., <0.5 percent TiO_2). They have ratios of more incompatible to less incompatible elements similar to some arc tholeiites (Figure 5). The extended REE patterns (Figure 6a) are distinctive with a flat LREE pattern at <1 x primitive mantle, a flat HREE pattern at ~1 x primitive mantle, and a distinctive 'step up' between Nd and Eu. These patterns closely resemble those of the Type A-III refractory island-arc tholeiites in the Wild Bight Group (Figure 6a). As noted above, such rocks are associated with most of the volcanogenic sulphide deposits in the Wild Bight Group and are known to occur in the footwall to the Point Leamington deposit.

Type II

Type II volcanic rocks are represented by a single sample, a subalkalic, low-TiO_2 (0.68 percent), LREE-depleted rock with characteristic arc-geochemical signature (Figure 6b). This rock has a Ti:V ratio characteristic of island-arc rocks (Figure 5) and is very similar, geochemically, to Type A-II LREE depleted arc tholeiites that occur in the Nanny Bag Lake area in the southern part of the Wild Bight Group. Such rocks in the Wild Bight Group occur in the intermediate part of the stratigraphy and are roughly stratigraphically equivalent to the refractory tholeiites and rhyolites.

Type III

Type III rocks are transitional tholeiitic to alkalic, high TiO_2 (~1.3 to 2 percent), LREE-enriched rocks. They have Ti:V ratios typical of rocks generated in non-arc environments and lack the arc-geochemical signature on the extended REE plots (Figure 6c). They are similar to 'transitional tholeiites' Type N-II) of oceanic island affinity that form part of the back-arc succession in the upper part of the Wild Bight Group.

There is a clear stratigraphic regularity to the occurrence of the geochemically different types of mafic volcanic rocks in the New Bay Pond sequence (Figure 2). The rocks of island-arc origin (Types I and II) occur only in the northeastern volcanic unit (i.e., adjacent to the West Arm Brook Thrust), whereas the non-arc (Type III) rocks are all in the southwestern volcanic unit. The geochemical data provide clear evidence that volcanic rocks in the New Bay Pond sequence have close analogues in the adjacent Wild Bight Group. In particular, the refractory arc tholeiites are a unique rock type that has previously only been recognized in the Wild Bight Group. These data strongly suggest that the northeastern volcanic belt in the New Bay Pond sequence can confidently be correlated with the central part of the Wild Bight Group and probably records part of the arc-rifting event. In contrast, the southeastern volcanic belt can be correlated with the upper parts of the Wild Bight Group, and records volcanism in the same back-arc basin (cf. Swinden *et al.*, 1990). The fact that most facing directions in the intermediate turbidite unit are toward the northeast suggests that there is further internal faulting within the sequence. It is herein informally proposed that rocks of the New Bay Pond sequence, as defined by Swinden (1988a), be included in the Wild Bight Group. Formal stratigraphic subdivision at the formation level awaits detailed mapping now in progress in central Notre Dame Bay.

GEOPHYSICS

The presence of refractory tholeiites in the New Bay Pond sequence that are almost identical to those found in close association with the Point Leamington deposit more or less along strike and only 7 km away, suggests that there may be a geological connection between the two areas. The strike extent of the refractory tholeiites and, in particular, their disposition between New Bay Pond and the Point Leamington deposit, is of obvious interest to explorationists. However, a straightforward along-strike correlation of favourable massive sulphide environments between the two areas is difficult to reconcile with previous interpretations of the local stratigraphy and structure that require a major discontinuity along the Long Pond Fault (Figure 2). Unfortunately, exposure is very poor in the critical area. In order to further investigate the possible geological connection between the Point Leamington deposit and the New Bay Pond area, an examination was made of: i) data from industry airborne electromagnetic (AEM) and ground electromagnetic surveys from the Geological Survey Branch assessment files; and ii) recent vertical gradient magnetic (gradiometer) surveys published by the Geological Survey of Canada.

ELECTROMAGNETIC SURVEYS

An AEM survey of the area between Big Lewis Lake and Seal Bay was carried out by Questor Surveys Limited (1972) using the INPUT system. Carbonaceous shales of the Shoal Arm Formation are highly conductive, and their surface extent is marked by prominent AEM anomalies that can be traced from northeast of New Bay Pond, northwestward across the western end of Big Lewis Lake to an area east of the Point Leamington deposit. The conductors correspond precisely to the mapped extent of the Shoal Arm Formation (Figure 2). The anomalies do not stop at the Long Pond Fault, as they should if this structure has a major offset, but appear to continue for some distance to the northwest of its mapped surface trace. More detailed electromagnetic information is provided by a Max-Min survey undertaken by Noranda Exploration Company Limited. The survey was carried out on 100-m lines and results are summarized in Figure 7, taken

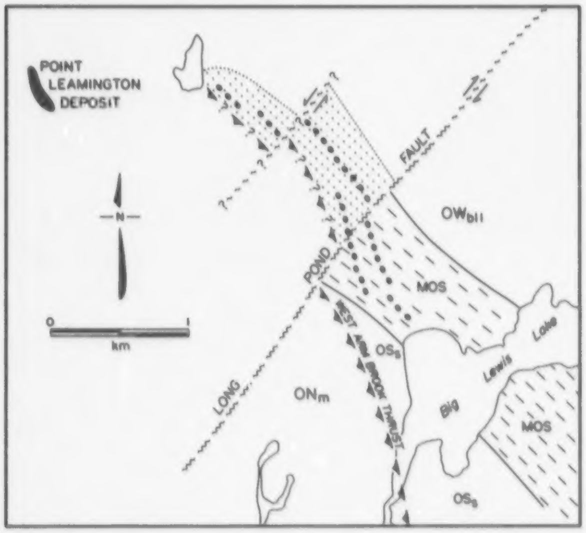


Figure 7. *Electromagnetic (Max-Min) anomalies (black ellipses) associated with carbonaceous shales of the Shoal Arm Formation northwest of Big Lewis Lake; legend as for Figure 2. The mapped extent of the Shoal Arm Formation and its previously supposed termination against the Long Pond Fault are shown by the dashed pattern. The Max-Min conductors strongly suggest the continuation of the Shoal Arm Formation northwest of the Long Pond Fault (dotted lines) with only minor dextral offset. The anomalies (and presumably the shales) terminate near a small pond east of the Point Leamington deposit.*

from a compilation map by Graves (1984). The EM anomalies on the Shoal Arm Formation clearly continue across the Long Pond Fault with only minor dextral offset. About 800 m to the northwest of the Long Pond Fault, they are again offset slightly in a sinistral sense and then terminate near a small pond, 1300 m east of the Point Leamington deposit.

The EM data lead to two principal conclusions. First, there is no geophysical evidence for a major offset along the Long Pond Fault; second, they strongly suggest that the Shoal Arm Formation shales (and presumably the West Arm Brook Thrust, which forms their western boundary in this area) can be traced continuously to a point where they are only 1300 m across strike from the Point Leamington deposit.

GRADIOMETER

Gradiometer data for the New Bay Pond area have recently been published by the Geological Survey of Canada (1988a,b) and show a good correspondence with known geology (Figure 8). Mafic volcanic and intrusive rocks are characterized by high magnetic relief, whereas sedimentary rocks generally have low relief.

The Big Lewis Lake basalt in the Wild Bight Group is a strongly magnetic unit marked by a strong, northwest-striking magnetic anomaly (Figure 8). Magnetic data indicate that this unit is slightly offset by several northeast-striking faults, including a sinistral offset on a previously unrecognized fault north of New Bay Pond and dextral offsets on the Four Mile Lake and Long Pond faults (consistent with EM data, see above). The Big Lewis Lake basalt can be traced continuously to the area northeast of the Point Leamington deposit where it forms part of a large, roughly elliptical 'bullseye' of high magnetic relief (Figure 8).

The Shoal Arm Formation and Sansom Greywacke overlying the Big Lewis Lake basalt have low magnetic relief, in sharp contrast to the adjacent volcanic rocks. The magnetic low that marks their outcrop extent can be traced continuously although with minor offsets along northeast-striking faults, to the area immediately east of the Point Leamington deposit. The disappearance of this magnetic low is coincident with the disappearance of the EM anomaly associated with the Shoal Arm Formation shales (Figures 7 and 8).

In the New Bay Pond sequence, the Type I and Type II mafic volcanic rocks in the northeastern volcanic belt are characterized by moderate magnetic relief, less pronounced than that associated with the Big Lewis Lake basalt. The fault contact between the volcanic rocks and the low-relief Sansom Greywacke and Shoal Arm Formation is sharply delineated. The linear magnetic high, associated with these volcanic rocks near New Bay Pond, can be traced northerly to the area of the Point Leamington deposit, where it converges and eventually merges with the Big Lewis Lake basalt anomaly to form the magnetic 'bullseye' northwest of Big Lewis Lake. Slight offsets along northeast-striking faults (e.g., the Four Mile Lake Fault) are consistent with offsets of other magnetic anomalies.

The turbidites in the central part of the New Bay Pond sequence are marked by generally lower magnetic relief, but contain two northwest-striking linear magnetic anomalies that are interpreted, on the basis of outcrop information, to reflect gabbroic sills (Figure 8). These anomalies are of some interest for the regional geological interpretations because they show the same dextral offset on the Four Mile Lake Fault as do the volcanic units, little or no offset on the Long Pond Fault and a minor dextral offset along the Seal Bay Brook Fault (Figure 8).

The Side Harbour volcanic unit is also marked by a prominent magnetic trend of moderate to high relief. Trending down from the north, it is slightly offset in a dextral sense along the Seal Bay Brook Fault (consistent with previous mapping); south of this fault, it abruptly terminates against a prominent, northeast-trending, magnetic low on the north side of the magnetic 'bullseye' that is transverse to the regional magnetic grain. Acrosss this magnetic low, the Side Harbour volcanic unit is approximately along strike with the Big Lewis Lake and New Bay Pond sequence basalts.

INTERPRETATION OF THE GEOLOGY BETWEEN NEW BAY POND AND THE POINT LEAMINGTON DEPOSIT

This paper proposes new interpretations of the nature and extent of the fault boundary between the New Bay Pond sequence and the Wild Bight Group–Shoal Arm Formation–

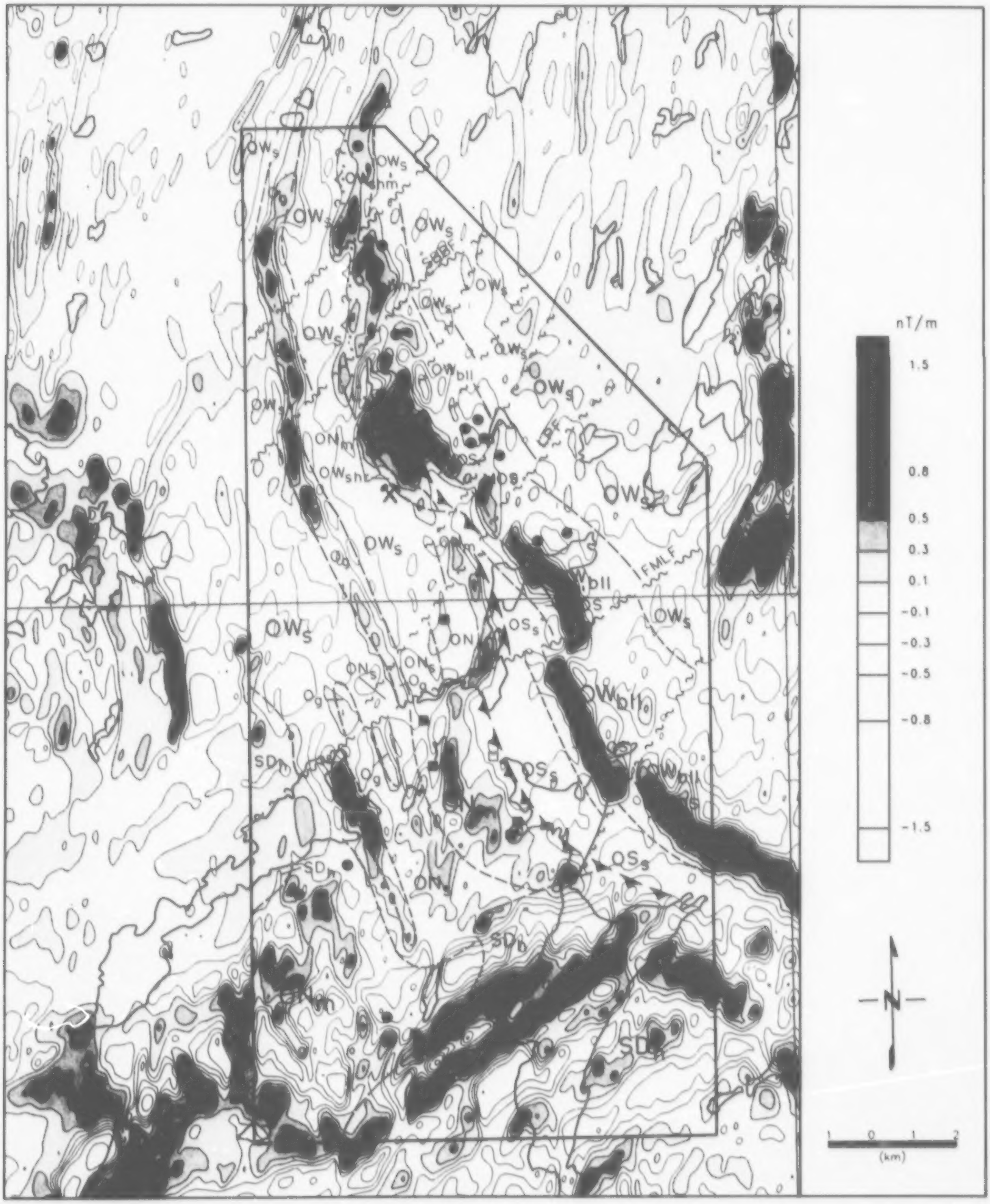

OWs
OWshm
SBBF
OWbll
ONm
OWshl
OS
OSs
Og
FMLF
ON
ONs
SDb
SDn
SDR
nT/m
1.5
0.8
0.5
0.3
0.1
-0.1
-0.3
-0.5
-0.8
-1.5
N
1 0 1 2
(km)

Figure 8. *Simplified gradiometer map for the same area as Figure 2 with geochemical information and newly interpreted geology in red. Legend and symbols as for Figure 2. Geochemical samples shown by squares and dots, as in Figure 5.*

Sansom Greywacke, the extent of volcanic rocks that are highly prospective for massive sulphides of the Point Leamington type, and the effect of late strike-slip faulting on stratigraphic reconstruction in central Notre Dame Bay.

THE WEST ARM BROOK THRUST

Kusky (1985) traced this fault from New Bay Pond northwestward to the north end of Big Lewis Lake. His observations led to two significant conclusions about its nature: i) it is a pre-folding thrust fault along which rocks of the New Bay Pond sequence were emplaced in a generally southerly direction over the Sansom Greywacke; and ii) in the area south of Big Lewis Lake, the fault ramps upward in the hanging wall, bringing New Bay Pond sequence rocks against increasingly higher stratigraphic levels of the Sansom Greywacke.

Data presented in this paper supports these conclusions and shows that they can be extended northwestward for some distance where they help to explain apparent anomalies in the succession in the area of the Point Leamington deposit. The ramping of the fault can be traced to the northwest (i.e., toward the source of the thrusting) on the gradiometer maps by the convergence of the high-relief anomalies associated with the mafic volcanic rocks at the top of the New Bay Pond sequence and the Big Lewis Lake basalt (Figure 8), with the concomitant thinning of the area of flat magnetic relief that marks the sediments above the Big Lewis Lake basalt. Although outcrop is extremely sparse northwest of Big Lewis Lake, the continued convergence of the two magnetic anomalies in this area indicates that structurally lower ramps originate in the Shoal Arm Formation and the New Bay Pond basalt is juxtaposed with successively lower carbonaceous shale strata toward the northwest. Immediately east of the Point Leamington deposit, the fault juxtaposes New Bay Pond sequence basalts with the Big Lewis Lake basalts. The West Arm Brook Thrust cannot be confidently traced north of the magnetic 'bullseye' formed by the structural juxtaposition of the two volcanic units. Detailed examination of shaded-relief maps of gradiometer data for this area (not shown) indicate that the fault continues in a northwesterly direction through the magnetic high (indicated by the extension of the West Arm Brook Thrust on Figure 8), and strikes into the transverse magnetic low that separates the 'bullseye' from the magnetic high over the Side Harbour volcanic unit.

RECOGNITION AND EXTENT OF MAFIC VOLCANIC UNITS

As previously recognized by Swinden *et al.* (1990), there are three mafic volcanic units in the immediate area of the Point Leamington deposit that can be distinguished on the basis of stratigraphic setting and geochemistry (Figure 8):

i) The Big Lewis Lake basalt—this consists of non-arc rocks of MORB and alkali basalt affinity. It is underlain by Wild Bight Group volcaniclastic turbidites but, except immediately northeast of the Point Leamington deposit, stratigraphically overlain by the Shoal Arm Formation. It is interpreted to lie at the top of the Wild Bight Group;

ii) The Side Harbour basalt (excluding rocks in the immediate footwall of the Point Leamington deposit)—this unit comprises a linear belt of transitional tholeiites and alkalic basalts of non-arc affinity. The volcanic rocks are underlain and overlain by Wild Bight Group volcaniclastic turbidites along contacts that have been considered to be stratigraphic (Dean, 1978; Swinden, 1987). On this basis, they have previously been interpreted to occupy a lower stratigraphic level in the Wild Bight Group than the Big Lewis Lake basalt;

iii) The refractory tholeiites in the Point Leamington deposit footwall are geochemically distinct from the Big Lewis Lake basalt and the rest of the Side Harbour volcanic unit, but identical to rocks that are interpreted to be stratigraphically lower in the Wild Bight Group then the Big Lewis Lake and Side Harbour mafic volcanic rocks.

Geochemical data show that refractory tholeiites in the footwall of the Point Leamington deposit are similar to those in the New Bay Pond sequence. Gradiometer data indicate that the volcanic unit is continuous between the two areas and can be traced with some precision, providing a good estimate of the extent of highly prospective ground south-southeast of the Point Leamington deposit. The recognition that this prospective volcanic unit sits immediately above an early thrust fault in the New Bay Pond area (the West Arm Brook Thrust), and the geophysical evidence that this fault continues northward to the area of the Point Leamington deposit, confirms previous suspicions (Swinden,1987) that the sequence hosting the Point Leamington deposit must be separated by a fault (or faults) from mafic volcanic rocks to the east and north.

The extent of prospective stratigraphy north of the Point Leamington deposit is still problematic, and interpretation of the extent of prospective stratigraphy north of the deposit depends in part on how one chooses to interpret the structure from geochemical and relatively inconclusive geophysical evidence. The West Arm Brook Thrust can be traced north of the deposit for about 1000 m; it eventually strikes into a northeast-striking linear, transverse magnetic low of uncertain origin (Figure 8). This magnetic low does not appear to represent a fault with major displacement as the gradiometer map shows, at most, only a slight sinistral offset on a mafic dyke about 2 km to the southwest along its trend. Therefore, it seems most likely that the West Arm Brook Thrust continues northward within or along the west side of the Side Harbour volcanic unit. This model is attractive because it provides a mechanism by which the along-strike, geochemically similar, and more or less geophysically continuous mafic volcanic rocks of the Side Harbour and Big Lewis Lake volcanic units can be be continuous with each other. In this model, the extensive turbidite sequence west of the Side Harbour unit would be structurally, not stratigraphically, above the volcanic rocks, thereby resolving the apparent dichotomy in the hanging-wall stratigraphy of the two volcanic units. Furthermore, it would obviate the necessity for a major structural dislocation transverse to stratigraphy in the area of the Point Leamington deposit that would otherwise be necessary to explain the apparently diverse stratigraphic relationships. The principal objection to this model is that no rocks with island-arc geochemical signatures have been identified north of the Point Leamington deposit on the west side of the Side Harbour volcanic unit (Figure 8). In fact, rocks immediately north of this feature and along strike with the Point Leamington deposit are geochemically characterized as transitional tholeiites of clear non-arc affinity (Swinden, 1987). Nonetheless, it is possible that the West Arm Brook Thrust continues in this direction, and that volcanic rocks with arc-geochemical signatures either are present and have not been sampled, have pinched out, or have been excised by another, as yet unrecognized thrust fault subparallel to the bedding. Clearly, if rocks with arc signatures are present west of the non-arc rocks of the Side Harbour volcanic unit north of the Point Leamington deposit, they would be highly prospective for base-metal exploration. Detailed geochemical studies along the western margin of the Side Harbour volcanic unit might go a long way toward resolving this question.

CONCLUSIONS

Application of new geological and whole-rock geochemical information, coupled with analysis of available gradiometer and electromagnetic data provide new interpretations of the geological relationships in the area between New Bay Pond and the Point Leamington massive sulphide deposit.

The New Bay Pond sequence contains two, northwest-striking volcanic belts. Geochemical data show that mafic volcanic rocks in the two belts are geochemically and tectonically distinct and correlate well with volcanic rocks in the adjacent Wild Bight Group. The northeasterly belt contains mafic volcanic rocks with island-arc geochemical signatures, including refractory tholeiites similar to those universally associated with volcanogenic sulphide deposits in the Wild Bight Group, and includes minor amounts of rhyolite. This unit can be traced northward by geochemistry and geophysics to the area of the Point Leamington deposit. The strike extent of this volcanic unit between the Point Leamington deposit and New Bay Pond is considered to be highly prospective for massive sulphide deposits, because it is along the strike extension of a horizon already demonstrated to be prospective and because it contains distinctive refractory basaltic rocks that are found to be closely associated with volcanogenic mineralization both in the Wild Bight Group and throughout central Newfoundland.

The southeastern volcanic belt of the New Bay Pond sequence has mafic volcanic rocks with non-arc geochemical signatures and is not considered to have potential for Point Leamington-type massive sulphide deposits.

The data show that a previously recognized early thrust fault (herein termed the West Arm Brook Thrust) separating the New Bay Pond sequence from the Sansom Greywacke can be traced northward to the area of the Point Leamington deposit. This provides good evidence that the Point Leamington deposit is structurally rather than stratigraphically above the non-arc rocks to the east, resolving a previously recognized stratigraphic enigma. The West Arm Brook Thrust (and hence, the prospective stratigraphy) can presently be traced with confidence for about 1 km northward from the Point Leamington deposit. However, geological modeling constrained by gradiometer data suggests that it may continue for some distance north of this along the western side of the Side Harbour volcanic unit. Detailed geochemical testing is required to determine whether volcanic rocks with arc signatures can be identified in the Side Harbour volcanic unit. Recognition of such rocks would provide evidence that the Point Leamington deposit stratigraphy continues to the north and would provide a good rationale for detailed base-metal exploration in this area.

The gradiometer data clearly show that the prominent set of northeast-striking faults in this part of Notre Dame Bay invariably have minor, variously dextral or sinistral, offsets. They do not appreciably disrupt the stratigraphy in the area between New Bay Pond and Seal Bay and cannot be called upon to explain apparent major discontinuities in the stratigraphy in this area.

ACKNOWLEDGMENTS

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GEOCHEMICAL MAPPING IN NEWFOUNDLAND AND LABRADOR: ITS ROLE IN ESTABLISHING GEOCHEMICAL BASELINES FOR THE MEASUREMENT OF ENVIRONMENTAL CHANGE

P.H. Davenport, T.K. Christopher[1], S. Vardy[2] and L.W. Nolan
Geochemistry and Geophysics Section

ABSTRACT

First-pass geochemical mapping of Newfoundland and Labrador was completed in 1987. For these geochemical surveys, organic sediment was collected from more than 40,000 lakes and the results used to promote mineral development by stimulating and focusing mineral exploration. These regional geochemical data are receiving increasing attention for the definition of geochemical baselines against which chemical change in the environment may be measured. To determine the natural geochemical background effectively for environmental purposes, it is necessary to demonstrate that the sample medium employed meets the following criteria: first, it is not already widely contaminated; second, that it reflects reliably the natural fluctuations in this background caused by geological variation; and thirdly, that it records environmental change where it has occurred.

Studies are in progress to evaluate organic lake sediment as sampled in the regional surveys against these criteria. The variation of trace-element levels in lake-sediment cores has been investigated in south-central Newfoundland to compare stratigraphic variance with spatial variance between lakes over geochemically contrasted geological units. Element-speciation studies employing partial and total digestions of the sediments indicate how trace elements are partitioned and hence their geochemical mobility and potential bio-availability. To determine the geochemical effects of urbanization, a lake-sediment survey was conducted of the St. John's metropolitan area.

From the results available, it is clear that lake sediment is a very appropriate medium for establishing geochemical baselines. In relatively pristine regions, the lake cores show that stratigraphic variation of trace-element concentrations in the upper metre of sediment, which are due to local environmental fluctuations over time, are generally quite minor and much smaller than geochemical variation between lakes due to the primary geochemical composition of the local bedrock. Significant proportions of a wide range of elements (i.e., Cu, Ni, Co, Zn, Y, Dy, La, Cd, Ce, Pb, Mn, V, Fe, Cr, Mo, Ba and Li) are held in the readily acid-leachable organic and amorphous oxide phases of the sediments, indicating their potential bio-availability under acid conditions. Finally, lake sediment from the St. John's region shows the effect of urbanization in a number of ways. This is most dramatically indicated by the increase in heavy metals such as Pb, which is from 5 to 10 times higher in lakes from the urban core than the normal upper limit of its natural background for the area.

INTRODUCTION

Man-made chemical changes to the environment are superimposed on a variable natural geochemical background where trace-element abundances range over several orders of magnitude. Since the chemical composition of the earth's surface directly affects the biosphere, a reliable overview of the natural variation in the geochemical background is essential in order to identify, measure and place into perspective the extent of man-made environmental changes. These changes may be local (e.g., the point discharge of pollutants to the environment or the modification of the natural geochemical background through activities such as construction), or widespread (e.g., airborne deposition of pollutants like acid rain).

The geochemical background at the earth's surface is determined by geology. The levels of the major elements (Si, Al, Fe, Mg, Ca, Na, and K) and their spatial variation can be estimated to a first approximation from geological and soil maps that are widely available. However, such maps offer very little information on the abundances of trace elements currently regarded either as essential to many forms of life (i.e., Cl, Co, Cu, F, I, Mn, Mo, P, S, Se, V and Zn) or as toxic (i.e., As, Be, Cd, Hg, Ni, Pb, Sb, Tl and U). Abundance levels for these important trace elements and their spatial variations can only be provided by appropriately designed geochemical surveys.

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[1] Department of Earth Sciences, Memorial University of Newfoundland

[2] Department of Geography, Memorial University of Newfoundland (currently Department of Earth Sciences, University of Waterloo)

Regional geochemical mapping in Newfoundland and most of Labrador is based on the composition of 'grab' samples of the upper 20 to 100 cm of organic sediment collected from about 40,000 lakes (Figure 1). The program commenced in 1973 following orientation studies that showed empirically that geochemical surveys of the Recent, organic, lake sediment was a cost-effective and rapid approach to mineral-resource appraisal and for focusing and stimulating mineral exploration in this province. Indeed, geochemical mapping using organic lake sediment has been widely and successfully applied in northern countries such as Canada (Friske and Hornbrook, 1991) and Finland (Tenhola, 1991), but studies to elucidate the geochemical processes that contribute to the success of the method have been relatively few (cf. Coker *et al.*, 1979).

Lake sediments reflect not only the spatial geochemical variations at the geosphere–biosphere interface but also, in many cases, they contain an historical record of the Holocene Epoch. Mineralogical and chemical changes may be preserved as a chemical stratigraphy (Engstrom and Wright, 1984). The bulk of an organic lake sediment is composed typically of indeterminate organic matter and fine-grained, inorganic material that is difficult to identify mineralogically. Chemical analysis provides the most practical way of quantifying compositional variation.

Chemical variation in a lake over time may be due to several factors that operate more or less contemporaneously. These factors include the intensity of erosion in the catchment, biological activity in the lake and variations in the groundwater regime. In addition, human activities may influence the chemistry of the upper sections of lake-sediment cores. The chemical record may be complex because of these manifold factors, and is best resolved with reference to other stratigraphic information from pollen assemblages, which provide information on climatic changes and anthropogenic activity, and diatom assemblages that reflect changes in lake-water alkalinity and salinity.

Chemical stratigraphy in lake sediment has been studied mainly to detect and characterize anthropogenic change around known sources of pollution such as mines (Qvarfort, 1983), smelters (Fortescue, 1986), and towns (Rogers and Ogden, 1991; Rogers *et al.*, 1991), as well as the effects of changes in land use (Burden *et al.*, 1986). These studies typically examine a few lakes in considerable detail with little reference to geologically determined geochemical variation. On the other hand, regional geochemical surveys for mineral-resource appraisal usually ignore or at least fail to quantify the nature and extent of anthropogenic changes that may mask the natural patterns of interest. A comparison of the upper 14 cm of lake sediment with sediment from below 20 cm was made from a collection of several hundred short cores from the Herman Lake map area north of Lake Superior in Ontario (Fortescue and Vida, 1991). The lower samples were deposited below the 'Ambrosia rise' (the layer at which a sharp increase in the proportion of pollen from ragweed occurs, which indicates the initial, widespread clearing of the forest in this area between 1880 and 1900), so by comparing the two sample sets, the nature and amount of environmental change in this century can be assessed. Increased Sb and As values in the post-Ambrosia sediments were noted in a fallout zone downwind from a sintering plant that has operated for over 50 years in Wawa, 20 km southwest of the study area. An area-wide increase in Pb values was also observed that showed no apparent relationship to the sintering plant fallout, whereas other elements such as Cu and Mn showed no overall difference between the two layers.

The study by Fortescue and Vida (1991) clearly indicates that an understanding of both spatial and temporal variations in the geochemistry of lake sediment is necessary to interpret the results of the existing regional geochemical database for the province from either an environmental or geological perspective. Two important questions are: 1) how variable are trace-element levels with depth in a typical lake-sediment profile and how does this stratigraphic variability compare with trace-element variability between lakes; and, 2) how sensitive are the 'grab' samples in detecting chemical changes due to human activities: can minor disturbances be safely ignored, and will major changes be detected?

To answer these questions, two studies have been conducted: the first (by Sheila Vardy) involved the collection and analysis of lake-sediment cores from five lakes in south-central Newfoundland from which the characteristics of chemical stratigraphy could be determined in regions little disturbed by human activity, and in the second (by Terry Christopher), 'grab' samples of organic lake sediment were collected throughout the St. John's urban area and environs to measure the effects of urban development.

METHODS

Lake-Sediment Cores

Sediment cores from five lakes in south-central Newfoundland (Figure 2) were collected in July 1989 (Vardy, 1990) primarily to study the history of deglaciation and the subsequent development of vegetation. Secondary objectives, and the aspects discussed here, were to study the chemical stratigraphy of the cores, both in relation to climatic change and the consequent changes in vegetation, and to determine the degree of variability in the upper 100 cm of sediment from which the 'grab' samples were obtained in the regional surveys. Although the main consideration in site selection was based on dating the deglaciation history, the north–south transect (Figure 2) crosses the main geological (and geochemical) trends at about 45°, so that significant differences in the levels of several trace elements were anticipated.

The cores were collected in duplicate from each lake using a Livingstone corer having a 5 cm internal diameter,

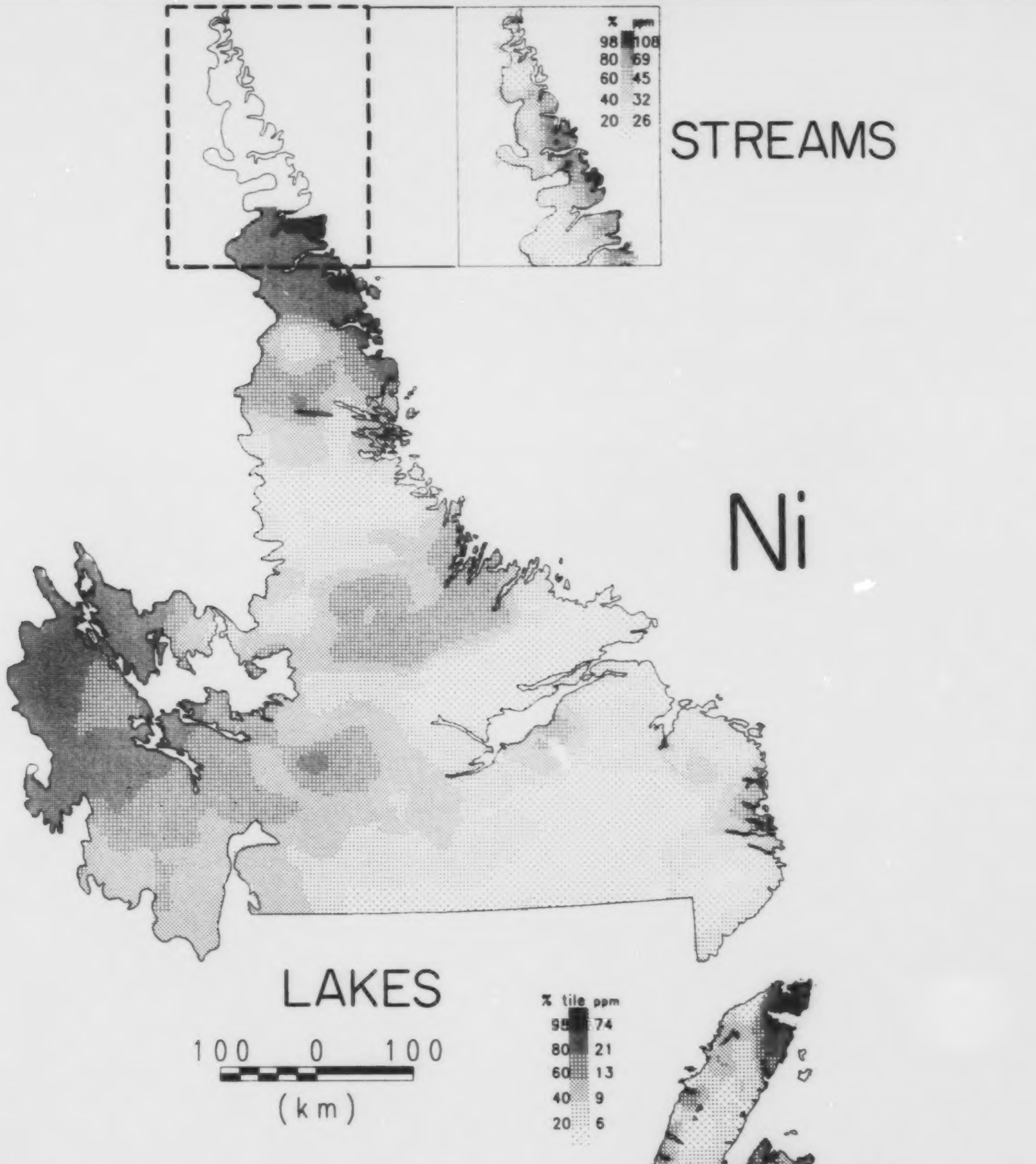

% ppm
98 108
80 69
60 45
40 32
20 26
STREAMS
Ni
LAKES
100 0 100
(km)
% tile ppm
98 74
80 21
60 13
40 9
20 6

Figure 1. *Nickel (Ni) distribution in drainage sediments in Newfoundland and Labrador.*

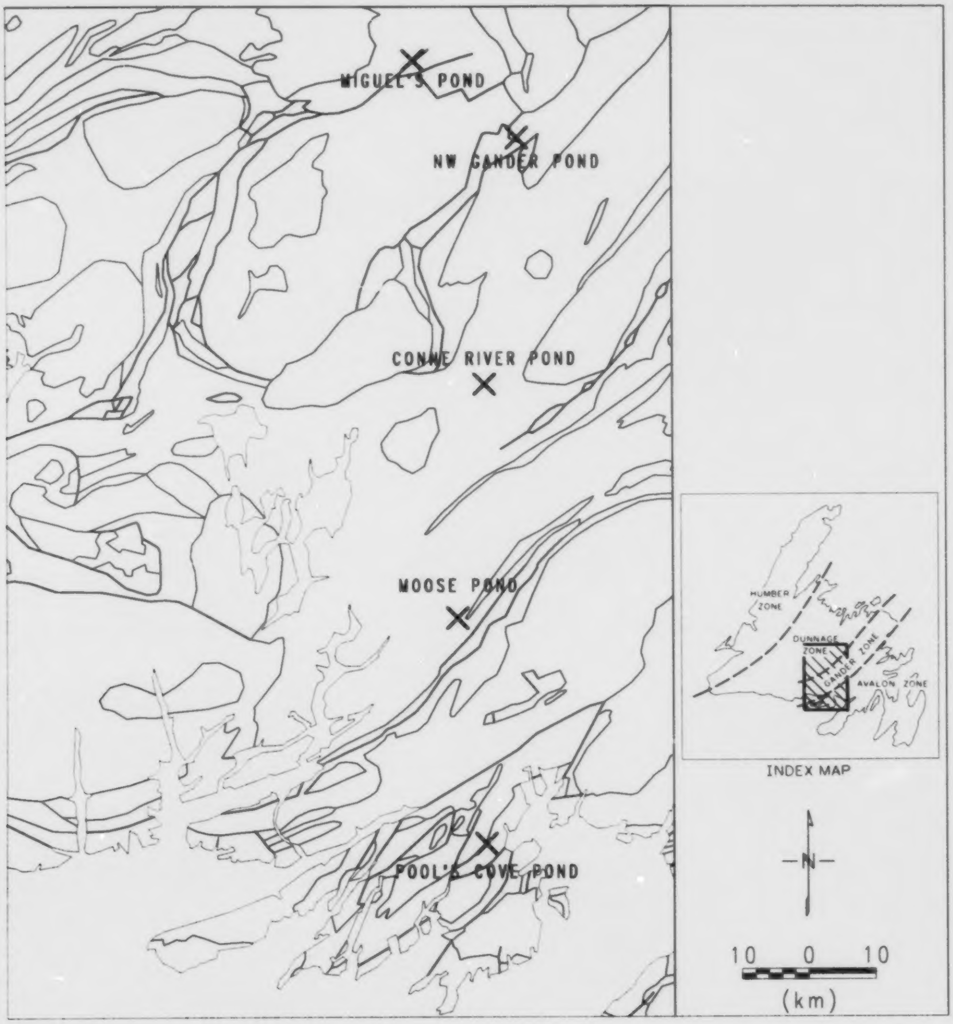


Figure 2. *The location of lakes from which sediment cores were collected in south-central Newfoundland; geological contacts from Colman-Sadd* et al. *(1990).*

and one of each pair was reserved exclusively for geochemical analysis. They were sampled at 5 cm intervals (except for the Pools Cove Pond core where, because of lack of material, the sampling interval was 25 cm). The samples were oven dried at 40°C, disaggregated in a porcelain mortar and pestle and passed through a 0.18 mm screen. Blind duplicate samples were introduced at this stage at a frequency of one pair in 20 samples to check analytical precision.

The dried (but unashed) sample material was used for analysis. Loss on ignition (LOI, a measure of organic content) was determined at 1000°C. The major elements (expressed as the oxides SiO_2, Al_2O_3, Fe_2O_3 (total Fe), MgO, CaO, K_2O, Na_2O, P_2O_5, TiO_2 and MnO), together with Ba, Cr, and Zr were determined by inductively coupled emission spectrometer (ICP-ES) following a lithium metaborate fusion (Wagenbauer *et al.*, 1983). Instrumental neutron activation analysis (INAA), was used to measure total concentrations of As, Au, Ba, Br, Ce, Co, Cr, Cs, Eu, Fe, Hf, La, Mo, Na, Ni, Rb, Sb, Sc, Se, Sm, Ta, Tb, Th, W, Yb and Zn. A combination of ICP-ES and atomic absorption spectrophotometry (AAS) was employed to determine total levels of Ba, Be, Cd, Ce, Co, Cr, Cu, Dy, Fe, Ga, La, Li, Mn, Mo, Nb, Ni, Pb, Rb, Sc, Y, Zn and Zr following an $HF\text{-}HClO_4\text{-}HCl$ digestion. The multi-element suites result in some data redundancy; there are three sets of results for Ba, Cr and Fe, and two for Ce, Co, La, Mo, Na, Ni, Rb, Sc, Zn and Zr. In most cases, there was a good correlation between results from the different methods, although for Ba, Cr, Ni, Zn and Zr the ICP-ES and AAS determinations were preferred because of the higher detection limits for these elements by INAA.

In addition, a 30 percent $H_2O_2-0.3M\ HCl$ digestion was used to extract elements held by organic matter and amorphous oxides (Engstrom and Wright, 1984), and the filtered extract was analysed by a combination of ICP-ES and AAS for Ba, Be, Cd, Ce, Co, Cr, Cu, Dy, Fe, Ga, La, Li, Mn, Mo, Nb, Ni, Pb, Rb, Sc, Y, Zn and Zr. The solid residue was digested in 0.2 M NaOH to dissolve biogenic (diatomaceous) silica (Engstrom and Wright, 1984), and this was determined by ICP-ES.

St. John's Urban Area Lake-Sediment Study

The method of sample collection in this study (Christopher, 1991) was the same as that used in the province-wide regional surveys in which 'grab' samples from the upper metre of organic lake sediment from the central, deeper basins of the lakes were obtained with a tubular steel gravity corer. A core 6 cm in diameter and 20 to 30 cm in length comprises a typical sample. As the sampler was allowed to free-fall from the water surface, the depth of penetration varied depending on the water depth and consistency of the sediment, and was observed to vary from as little as 20 cm to as much as 1[illegible] cm. Organic lake sediment was collected from most lakes within a 320 km² area that included the urban core and suburbs (Figure 3). Duplicate samples were taken from 52 of the 60 sample sites to allow sampling variance to be well characterized. The sediments were oven-dried at 40°C, mechanically crushed and sieved to <0.18 mm to remove any coarse debris and homogenize the material. The dried (but unashed) sediments were analyzed for 47 elements (total content): As, Au, Br, Co, Cs, Eu, Fe, Hf, La, Lu, Mo, Na, Sb, Se, Sm, Ta, Tb, Th, U, W and Yb by INAA, and Al, Ba, Be, Bi, Ca, Ce, Cr, Cu, Dy, Ga, K, Li, Mg, Mn, Nb, Ni, P, Pb, Rb, Sc, Sr, Ti, V, Y, Zn and Zr by a combination of ICP-ES and AAS, following an HF-$HClO_4$-HCl digestion. Organic carbon content was estimated from loss on ignition at 500°C and, in addition, the National Geochemical Reconnaissance (NGR) *aqua-regia* digestion (4M HNO_3-1 M HCl) was used to determine acid-soluble levels of Ag, Cd, Co, Cu, Fe, Mn, Ni, Pb and Zn by AAS (Friske and Hornbrook, 1991). Comparison of the total and *aqua-regia* extractable values showed the latter to be very nearly total for this whole suite of elements. These analytical methods are the same as those employed in the regional surveys.

Comparison of Analytical Methods

The methods used for total-element analysis in the lake-core study and the St. John's and regional surveys are directly comparable. Two different partial analytical methods were used and the 'efficiency' of these extractions was estimated by comparing the median values of the partial and total digestions for each element and their degree of correlation on a sample-by-sample basis (Table 1).

Considering first the H_2O_2-HCl partial digestion, most elements show a significant correlation between total and partial data (Table 1), which indicates that they are present mainly in the organic or amorphous oxide phases that are leachable in dilute acid. In general, those elements where the H_2O_2-HCl digestion is closest to total show the greatest correlation between partial and total data. Exceptions are Cd, where the rather low correlation is due to relatively large analytical errors at the low levels present, and Li, Fe and Mn, where the correlations are very high, but the proportion of metal extracted is small (17 to 47 percent). There was no statistical correlation between the results from the partial attack for Ga, Nb, Rb, Sc, Th or Zr and the total element values, with the partial results being much lower and showing little overall variation. For Rb, Sc and Zr, this may indicate

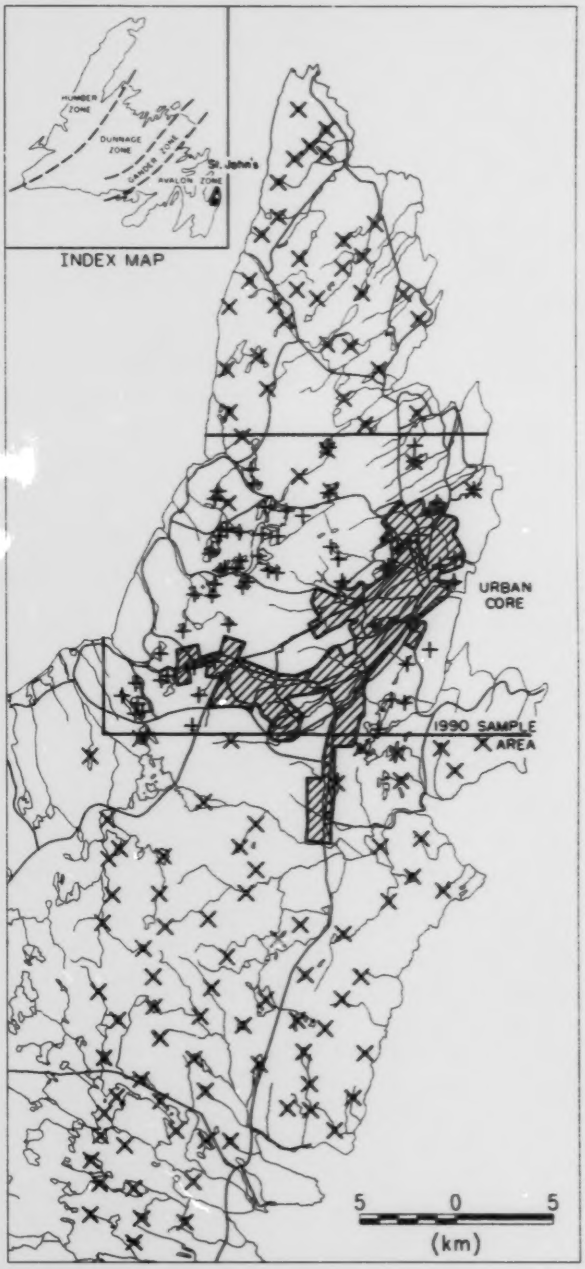


Figure 3. *The location of lakes sampled in the St. John's urban area (+ symbols) in relation to the province-wide lake-sediment sample sites (X symbols).*

Table 1. Trace-element leaching tests: the efficiency of 30 percent $H_2O_2-0.3M\ HCl$ and $4M\ HNO_3-1M\ HCl$ digestions on organic lake sediment relative to total (HF-$HClO_4$-HCl or INAA) data. Elements are listed in decreasing order of 'extraction' as measured by the ratio of partial to total values. There was no significant fraction of Ga, Rb, Sc, Sr, Th or Zr extracted by the 30 percent $H_20_2-0.3M\ HCl$ digestion

	30 percent $H_2O_2-3M\ HCl$		$4M\ HNO_3-1M\ HCl$	
Element	median ratio^0	correlation^2	median ratio^1	correlation^2
Cu	1.00	0.97	1.00	0.99
Ni	0.98	0.98	1.00	0.96
Co	0.87	0.97	0.95	0.99
Zn	0.85	0.94	0.89	0.98
Y	0.78	0.81		
Dy	0.71	0.91		
La	0.67	0.73		
Cd	0.67	0.58		
Ce	0.56	0.75		
Pb	0.50	0.69	0.76	0.96
Mn	0.47	0.98	1.00	0.98
V	0.38	0.56		
Fe	0.34	0.93	1.00	0.99
Cr	0.30	0.67		
Mo	0.29	0.81		
Be	0.24	0.72	0.36	0.87
Ba	0.24	0.67		
Li	0.17	0.97		

[1] ratio of the median value from 'partial' digestion to the median for the 'total' digestion

[2] degree of linearity between 'partial' and 'total' data as expressed by the product-moment correlation coefficient.

that these elements are present mainly in the silicate phase of the sediment, which is not dissolved in the partial digestion. For Ga, Nb and Th, the data are close to the analytical detection limit and any real correlations would be obscured by analytical noise.

Data to estimate the efficiency of the NGR ($4M\ HNO_3-1\ M\ HCl$) digestion for lake sediment are available from the St. John's study area only for Cu, Ni, Co, Zn, Pb, Mn and Fe (Table 1). The data for Be is for samples from the Strange Lake area of Labrador (J.W. McConnell, personal communication, 1991). A rigorous comparison between the two partial digestions is not possible because different sample sets are involved, but the results do suggest that the NGR digestion is total for Cu, Ni, Mn and Fe, and nearly so for Co and Zn, and is more efficient than the 30 percent $H_2O_2-0.3\ M\ HCl$ digestion for Pb and Be.

CHEMICAL STRATIGRAPHY IN LAKE SEDIMENT IN CENTRAL NEWFOUNDLAND

Geochemical profiles for As and Sb (total content) in cores from the five lakes are plotted next to the regional distribution maps in Figures 4 and 5. Although some variability of element concentrations with depth is apparent, the upper metre or so is generally quite consistent in composition and the differences in element abundances between lakes predicted by the regional survey data are reliably reproduced throughout the upper parts of the lake-sediment cores. Indeed, close correspondence between values based on single grab samples from the regional survey and the median value in the upper 120 cm of the sediment profiles is observed (Table 2) for most elements determined for the three lakes sampled in both studies (i.e., Miguel's Pond, Northwest Gander River Pond and Moose Pond). The Conne River site and Pool's Cove Pond were not sampled in the regional survey, so data from the closest sampled lake is shown in Table 2 and, as might be expected, the correspondence between interpolated regional and profile data is not as close. Considering the regional and profile data together, the areal variations in As and Sb, which are related to the regional geology, are much larger than within lake variation. The same is true for most trace elements evaluated.

The levels of H_2O_2-HCl extractable Pb show a marked increase in the uppermost 10 to 20 cm in two of the cores (Miguel's Pond and Moose Pond) and, less obviously, in Pool's Cove Pond and the Conne River site (Figure 6). These trends appear similar to those described from lake sediment in Ontario (Fortescue and Vida, 1991). Brunskill *et al.* (1991) describe a build up of Pb in lake sediments that were deposited since 1900 from a transect in central Canada between 49°N and 82°N, in which the amount of excess Pb decreases from south to north. Brunskill *et al.* (1991) attributed this pattern to long-range atmospheric transport of pollution, which can apparently be detected thousands of kilometres from source. The data reported here should be regarded as preliminary as the particular acid extraction used is not ideal for Pb because of the low solubility of $PbCl_2$. This trend is much less obvious in total Pb.

The largest variations in the concentrations of some elements are observed in the lower core sections. Core lengths recovered vary from 200 cm in Northwest Gander River Pond to 550 cm in Miguel's Pond, and in all lakes except Miguel's Pond the transition between organic sediment or gyttja and the underlying clastic sediments was penetrated (Vardy, 1990). This change in mineralogical composition is marked by large changes in the levels of LOI, biogenic silica, major elements and some trace elements. The abrupt increases in level of As, Sb and Pb at depths greater than 100 cm in Northwest Gander River Pond occur at this sediment transition. Other large changes in trace-element concentrations are present just above this transition, and may be geochemical reflections of climate change and related changes in erosional and vegetative regimes analogous to those described by Fortescue (1986) in Ontario and Rogers *et al.* (1991) in Nova Scotia. Some radiocarbon dating of samples from these transitional sections is

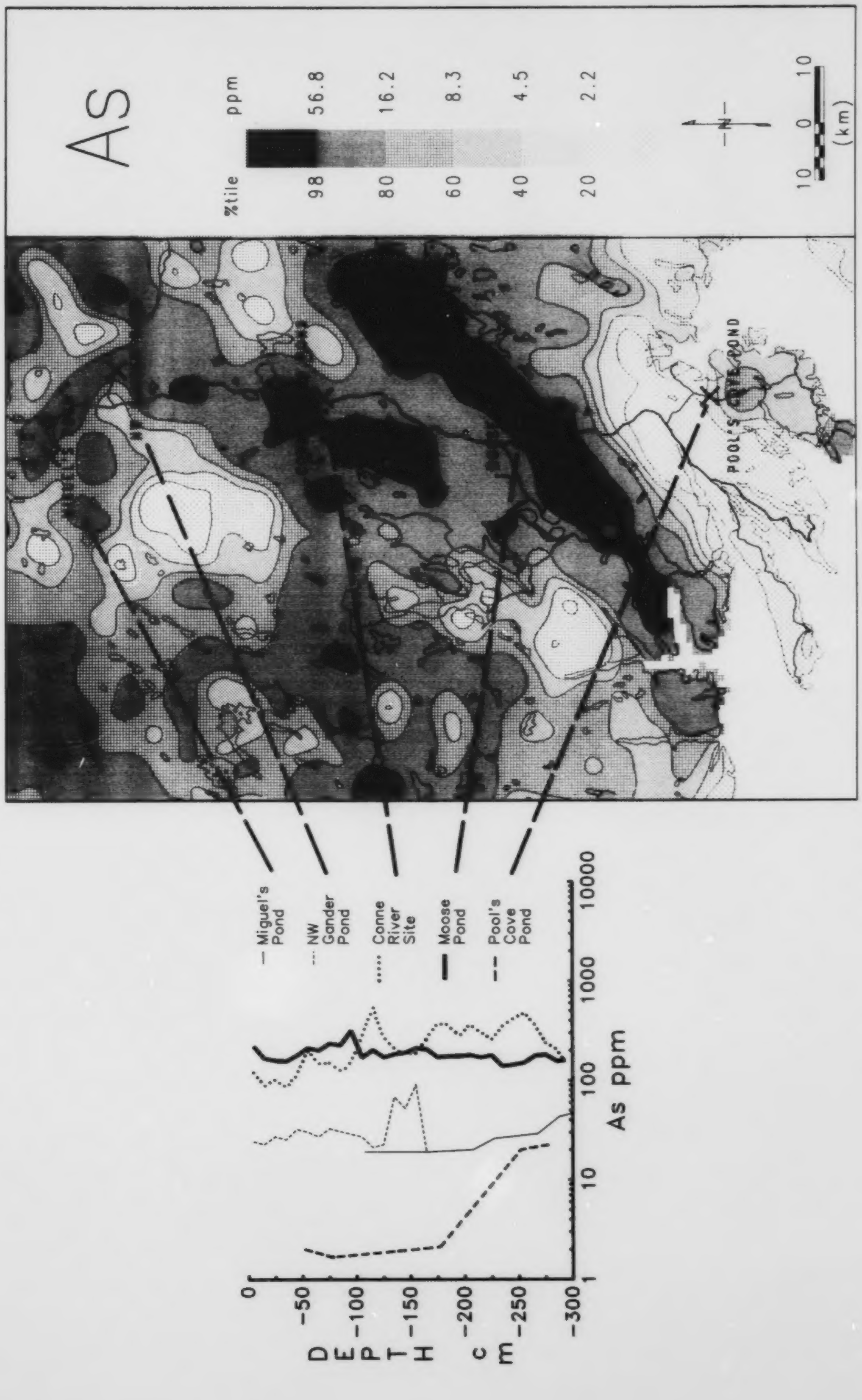

As
%tile
ppm
98
56.8
80
16.2
60
8.3
40
4.5
20
2.2
N
10
0
10
(km)
POOL'S COVE POND
Miguel's Pond
NW Gander Pond
Conne River Site
Moose Pond
Pool's Cove Pond
DEPTH cm
0
-50
-100
-150
-200
-250
-300
1
10
100
1000
10000
As ppm

Figure 4. *Arsenic (As) in lake sediment: its areal distribution (right) and (left) its distribution with sediment depth (or time).*

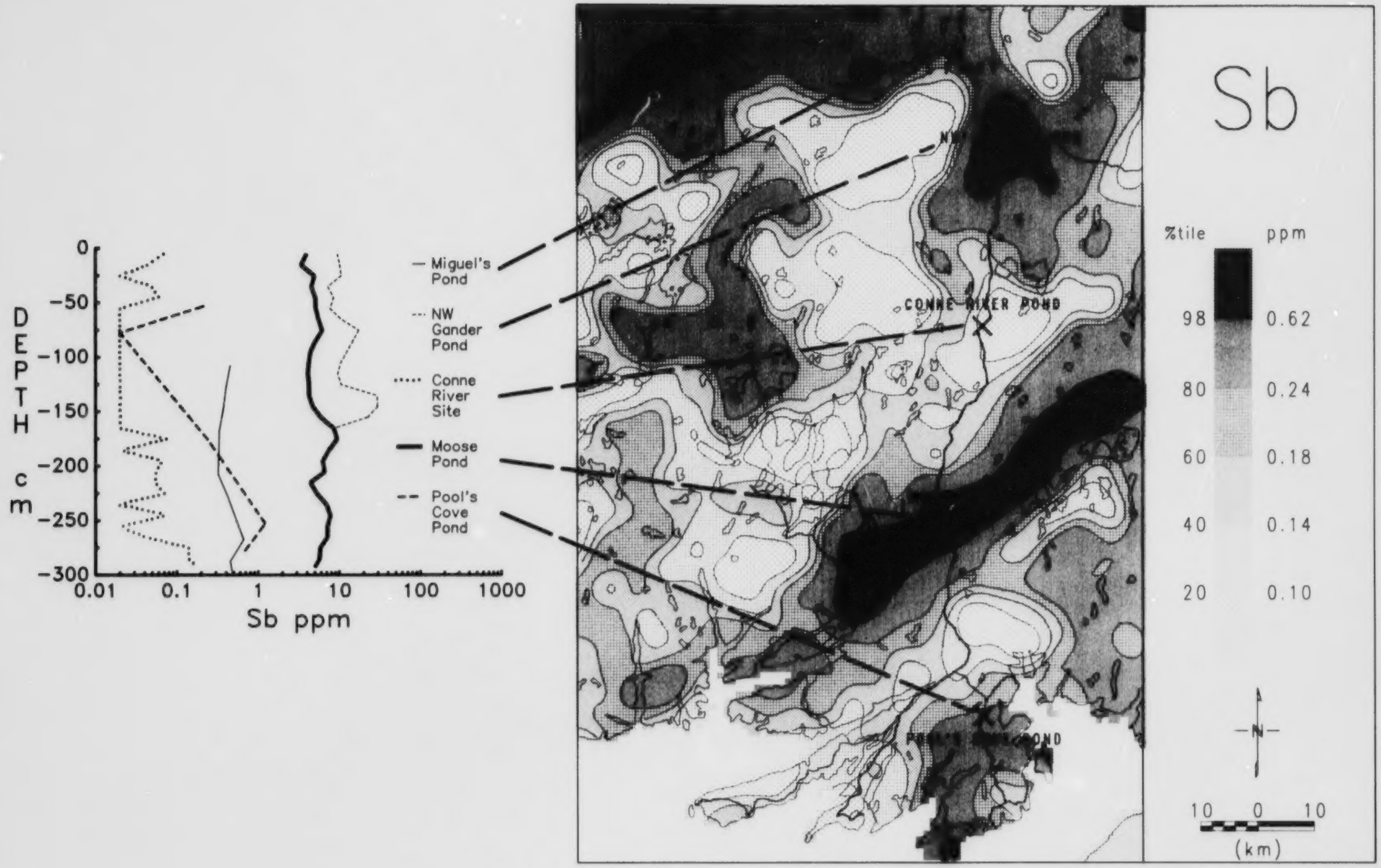

DEPTH cm
0
-50
-100
-150
-200
-250
-300
0.01
0.1
1
10
100
1000
Sb ppm
Miguel's Pond
NW Gander Pond
Conne River Site
Moose Pond
Pool's Cove Pond
NW
CONNE RIVER POND
POOL'S COVE POND
Sb
%tile
ppm
98
80
60
40
20
0.62
0.24
0.18
0.14
0.10
N
10
0
10
(km)

Figure 5. *Antimony (Sb) in lake sediment: its areal distribution (right) and (left) its distribution with sediment depth (or time).*

Table 2. Comparison of geochemical data from 'grab' samples of lake sediment collected in regional geochemical mapping with the median values from the upper 120 cm of lake-sediment cores sampled at 5 to 25 cm intervals. All data for As, Ba, Cr, Sb, U and W are total values: data for Co, Cu, Mn, Ni, Pb and Zn in 'grab' samples from the regional survey are from NGR *aqua-regia* digestions; for the cores, medians of these elements from both the H_2O_2−0.3M HCl digestion (Part. column) and total values (Tot. column) are given. Regional values are not available for the Conne River site and Pool's Cove Pond; the values in parentheses are from the nearest sampled lakes

	Miguel's Pond			NW Gander R. Pond			Moose Pond			Conne River Site			Pool's Cove Pond		
	Grab	Core		Grab	Core		Grab	Core		Grab	Core		Grab	Core	
		Part.	Tot.		Part.	Tot.		Part.	Tot.		Part.	Tot.		Part.	Tot.
As (ppm)	29.5	–	19.0	20.2	–	23.0	145	–	198	(16.0)	–	128	(1.7)	–	2.0
Ba (ppm)	<50	–	<50	120	–	100	520	–	495	(56)	–	83	(220)	–	<50
Co (ppm)	8	6.5	9	22	18	20	42	47	58	(8.5)	13.0	13.5	(5)	3	2
Cr (ppm)	21	–	23	190	–	129	42	–	52	(<15)	–	36	(16)	–	20
Cu (ppm)	12	19	18	15	19	20	19	19	19	(9)	14	12	(17)	11	12
Mn (ppm)	693	400	300	252	200	300	3340	3400	5000	(357)	700	1300	(133)	100	100
Mo (ppm)	4	–	3	2	–	3.0	7	–	9	(2.5)	–	6.5	(6)	–	30
Ni (ppm)	32	35	40	172	184	264	36	41	49	(11.5)	21	24	(12)	4	18
Pb (ppm)	11	8	12	5	6	18	19	13	18	(2.5)	5.0	13.5	(18)	11	17
Sb (ppm)	0.24	–	0.45	7.81	–	6.2	5.23	–	4.5	(0.05)	–	<0.05	(0.22)	–	0.23
U (ppm)	2.1	–	2.4	1.2	–	1.2	2.4	–	3.0	(1.0)	–	2.1	(5.0)	–	15.0
W (ppm)	<2	–	<2	<2	–	<2	<2	–	2.2	(13.0)	–	19.0	(3.4)	–	8.0
Zn (ppm)	24	282	284	59	100	127	143	193	228	(29.5)	104	110	(29.5)	78	90

discussed by Vardy (1991), but a full interpretation of the geochemical stratigraphy will require further palynological studies. These major features of the geochemical stratigraphy are well below the sediment layer sampled in the regional surveys in this area.

MAPPING GEOCHEMICAL CHANGES IN URBAN DRAINAGE SYSTEMS USING ORGANIC LAKE SEDIMENTS

Natural variations in the geochemical background for Pb are displayed in the regional lake-sediment data (Figure 7, left), which can be linked to broad-scale variations in the Pb content of bedrock, such as elevated values in belts of posttectonic granites and widespread Pb mineralization in the Cambro-Ordovician carbonate sequences of western Newfoundland. The Avalon tectono-stratigraphic zone in southeastern Newfoundland has a higher regional background for Pb than the central part of the Island (Dunnage zone), and within the northeastern part of the Avalon Peninsula the geological control of background Pb values is displayed at a more local level by elevated Pb in lake sediment over the shale-rich St. John's Group (HSJ; Figure 7, right). The highest Pb values are concentrated where the urban area overlies the St. John's Group, and there is a clear clustering of the highest values over the city core and the older suburbs. Peak values are up to 10 times higher than the natural background for the area, and above the 99th percentile for all of Newfoundland.

For Sb, lake-sediment values in the Avalon Zone are lower than in the eastern part of the Dunnage Zone, and fall between the 20 and 80th percentiles for the whole Island (Figure 8). Within the northeast Avalon area there is a marked clustering of high Sb values in lakes in the urban core, suggesting the possibility of anthropogenic input. Peak values of over 0.6 ppm Sb in lake sediment in St. John's are above the 92nd percentile for the Island as a whole. For As, the regional picture is similar (Figure 9), with the eastern part of the Dunnage Zone showing a broad pattern of enrichment; values in lake sediment on the northeast part of the Avalon Peninsula fall in the middle of the natural range for the Island as a whole. In contrast to Sb, however, As values in lake sediments from the urban core are close to the median of 6.2 ppm for all of Newfoundland.

CONCLUSIONS

The two detailed studies discussed here provide insight into the use of the data from regional geochemical mapping in establishing reliable and relevant geochemical baselines against which environmental change can be measured and its significance assessed. The study of chemical stratigraphy confirms that, throughout most of Newfoundland and for most elements, a 'grab' sample taken from within the upper metre of the profundal lake-sediment section provides data representative of the local geochemical background. The main cause of geochemical variation is geological, and anomalies in undeveloped regions may be expected to be both reproducible, and related to real geological features such as mineralization.

The geochemical patterns in lake sediment within areas of fairly intensive and sustained development, such as in the St. John's urban region, show, however, that pollution can completely obscure natural patterns for some elements. Taken

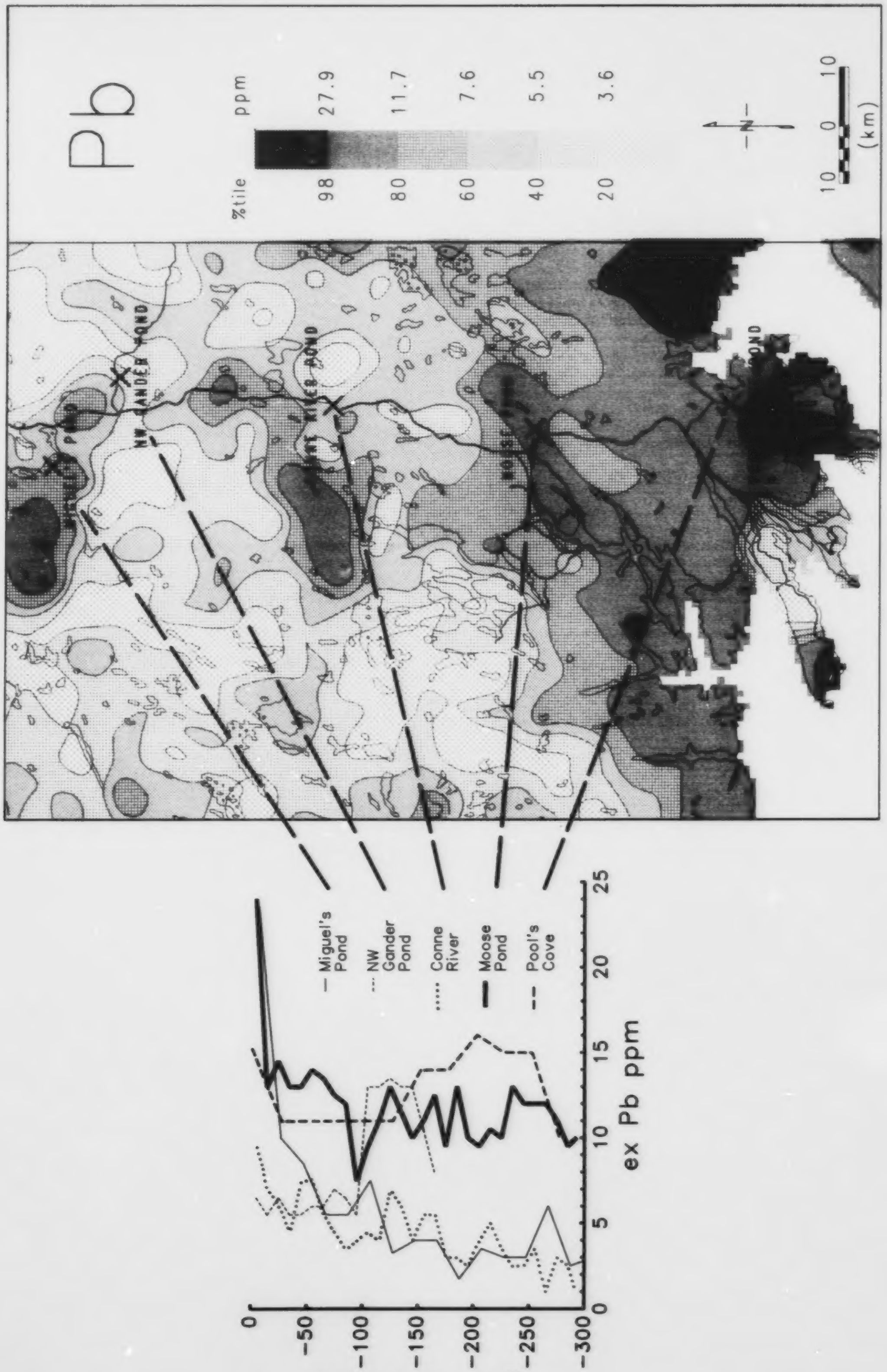
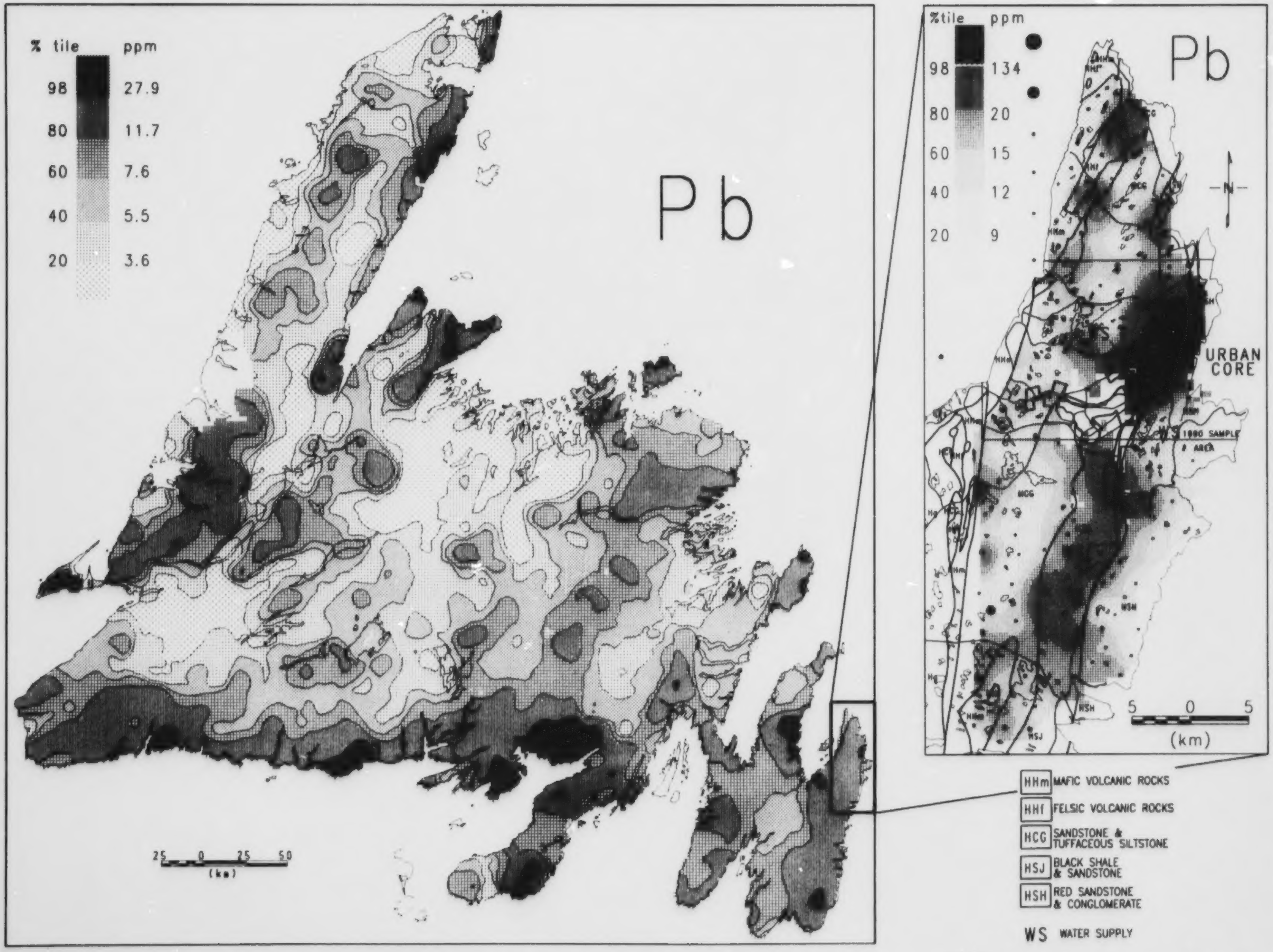

Pb
%tile
ppm
98
27.9
80
11.7
60
7.6
40
5.5
20
3.6
N
10
0
10
(km)
MIGUEL'S POND
NW GANDER POND
CONNE RIVER POND
MOOSE POND
POND
Miguel's Pond
NW Gander Pond
Conne River
Moose Pond
Pool's Cove
DEPTH cm
0
-50
-100
-150
-200
-250
-300
ex Pb ppm
0
5
10
15
20
25

Figure 6. *Lead (Pb) in lake sediment: its areal distribution (right) and (left) its distribution with sediment depth (or time).*


% tile
ppm
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27.9
80
11.7
60
7.6
40
5.5
20
3.6
Pb
25 0 25 50
(km)
%tile
ppm
98
134
80
20
60
15
40
12
20
9
Pb
N
URBAN CORE
WS
1990 SAMPLE AREA
5 0 5
(km)
HHm MAFIC VOLCANIC ROCKS
HHf FELSIC VOLCANIC ROCKS
HCG SANDSTONE & TUFFACEOUS SILTSTONE
HSJ BLACK SHALE & SANDSTONE
HSH RED SANDSTONE & CONGLOMERATE
WS WATER SUPPLY

Figure 7. *Regional distribution of Pb in lake sediment from 16,000 sites throughout Newfoundland (left) and, in more detail on the right, from 246 sites on the northeast Avalon Peninsula, including the St. John's urban region. Regional geology is from King (1988).*

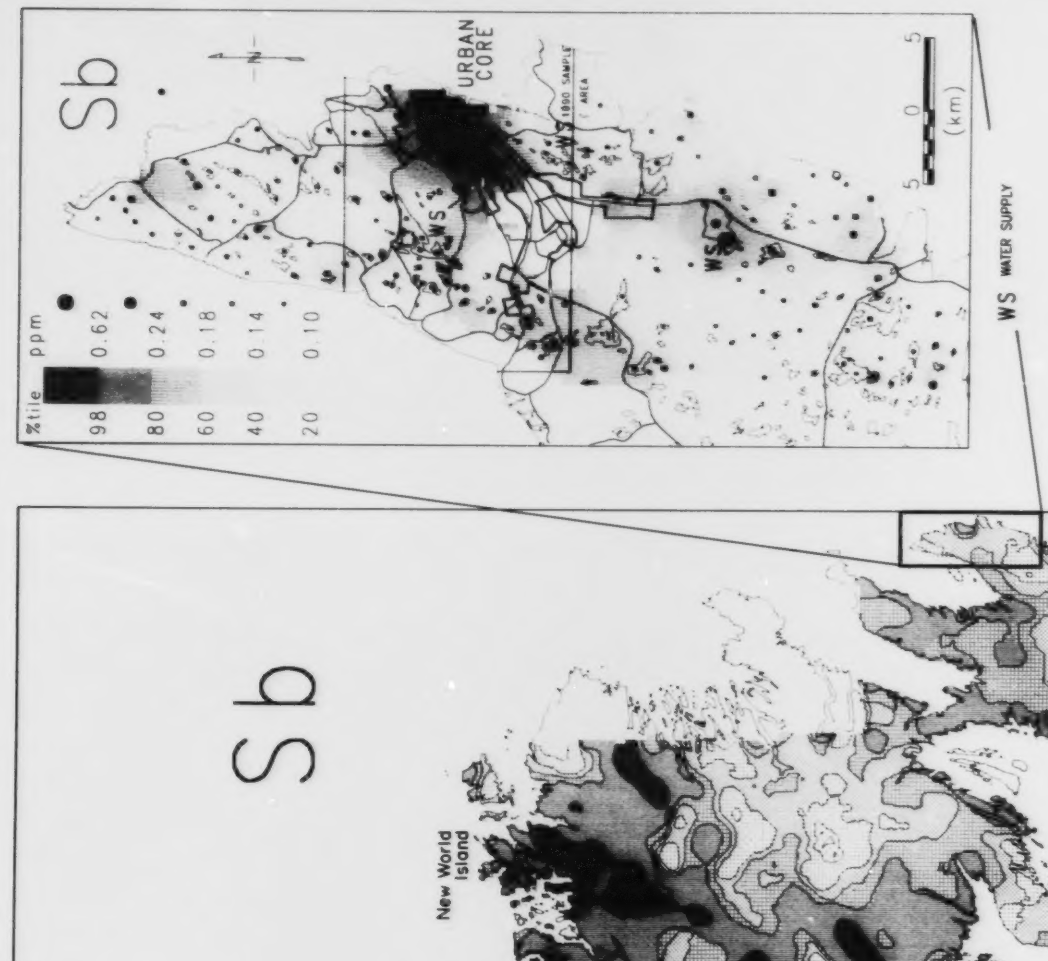
Sb
% tile ppm
98 0.62
80 0.24
60 0.18
40 0.14
20 0.10
New World Island
Bay D'Espoir
25 0 25 50
(km)
Sb
% tile ppm
98 0.62
80 0.24
60 0.18
40 0.14
20 0.10
N
URBAN CORE
1990 SAMPLE AREA
WS
5 0 5
(km)
WS WATER SUPPLY

Figure 8. *Regional distribution of Sb in lake sediment throughout Newfoundland (left) and, in more detail on the right, from 246 sites on the northeast Avalon Peninsula, including the St. John's urban region.*

together, these results imply that for mineral exploration the sampling strategy employed in the regional surveys, both provincially and nationally, is sound and reliable except in obviously disturbed areas, where other approaches should be adopted. These might include sampling the lake sediment below the contaminated layer (Fortescue and Vida, 1991) or employing a sample medium unaffected by pollution.

The apparent enrichment in Pb in the uppermost sections of some of the cores in central Newfoundland suggests that the region may be affected by airborne pollution. More work is required to confirm these patterns, however, and to document their origin, if they are indeed real. Their restriction to the uppermost sediment layers and their relatively modest intensity suggest that they will not substantially affect the measurement of the natural Pb distribution patterns. It may be desirable, however, to discard the upper 10 to 20 cm of sediment in geochemical exploration surveys as a precautionary measure.

A comparison of the weak acid digestions with the total-element determinations (Table 1) permits some preliminary conclusions about the partitioning of the elements within the different phases of the lake sediment and about element mobilities and their potential bio-availability. First, both Cu and Ni are strongly partitioned into the organic phase and are completely leached by both partial digestions (30 percent H_2O_2−0.3 M HCl and 4M HNO_3−1M HCl). This conclusion is supported by the generally observed correlation in the regional lake-sediment data between these two elements and organic content. Second, the greater efficiency of the NGR digestion in dissolving Mn and Fe oxides and hydroxides is probably responsible for the higher proportions of Co, Zn and Be extracted, and suggests that these oxide-hydroxide phases, together with the organic phase, contain a significant fraction of these elements. This conclusion is also in accord with observed correlations for the regional dataset as a whole. The ability of dilute, oxidizing acid digestions to leach a significant and fairly consistent proportion of the majority of elements determined suggests these elements possess a good deal of geochemical mobility in the surficial environment, and indicates that these partial digestions provide a fairly reliable measure of variations in concentration of elements potentially available to biological systems. Even elements such as Ba, Pb and Cr, normally considered immobile under normal surface conditions from thermodynamic data (Brookins, 1988), clearly have a labile fraction in lake sediment. These observations suggest that geochemical data based on dilute acid leaches of lake sediment (such as those employed in Canada's National Reconnaissance Program) can establish geochemical baselines that are very relevant to the measurement of environmental change.

The results from the St. John's area illustrate the relevance to environmental studies of the regional geochemical data based on organic lake sediment. The medium is clearly sensitive to environmental change, both natural and anthropogenic and, in order to assess the significance of anthropogenic change, it is essential to establish the bounds of natural variation. Lakes in the oldest and most urbanized part of St. John's show considerable geochemical change with elevated levels in sediment of many metals (Ag, Au, Ba, Cd, Cr, Cs, Cu, F, Hf, Lu, Ni, Rb, Sc, Ta, Th, U, V, W, Yb and Zn) in the inner city core (Christopher, 1991), but not in all cases above geochemical background. Levels of Pb in lake sediment from the urban core are above the 99th percentile of values found in all Newfoundland; natural values at this level result from Pb-bearing mineral deposits in the catchment but no Pb mineralization is known in the St. John's area. In contrast, although Sb shows a well-defined pattern of enrichment in lakes in the urban core, this enrichment is relative to the local background only. There are extensive areas between New World Island and Bay d'Espoir (Figure 8) where Sb values in lake sediment are as high or higher; this is thought to be due to the widespread effects of major hydrothermal systems (Davenport and Nolan, 1991). As a third example, for As there is no evidence of any enrichment pattern in the urban area and, in fact, level. of As on the northeastern Avalon Peninsula are normal relative to most of Newfoundland. Potential environmental problems with Sb and As are more likely to result should construction and development occur in areas naturally enriched in these elements, as happened in Nova Scotia in the vicinity of Halifax airport (Guilcher, 1987; Worgan, 1987), rather than in the St. John's region.

Lake sediments give a geochemical signal integrated over several years of deposition that reflects the sum of human activities (farming, deforestation, construction, waste disposal, industry, etc.) as well as natural physical, chemical and biological processes operating in the catchment. Where human disturbance has been intensive and sustained over several decades the geochemical changes are obvious even in a quick reconnaissance survey in which 'grab' samples of sediment are collected. If adequate baseline data are available, this approach provides a cost-effective method of gaining an overview of the nature and location of environmental changes, which allows follow-up to be focused on the main problem areas through more careful study of geochemical stratigraphy to determine a detailed history of change (e.g., Qvarfort, 1983; Rogers *et al.*, 1991).

ACKNOWLEDGMENTS

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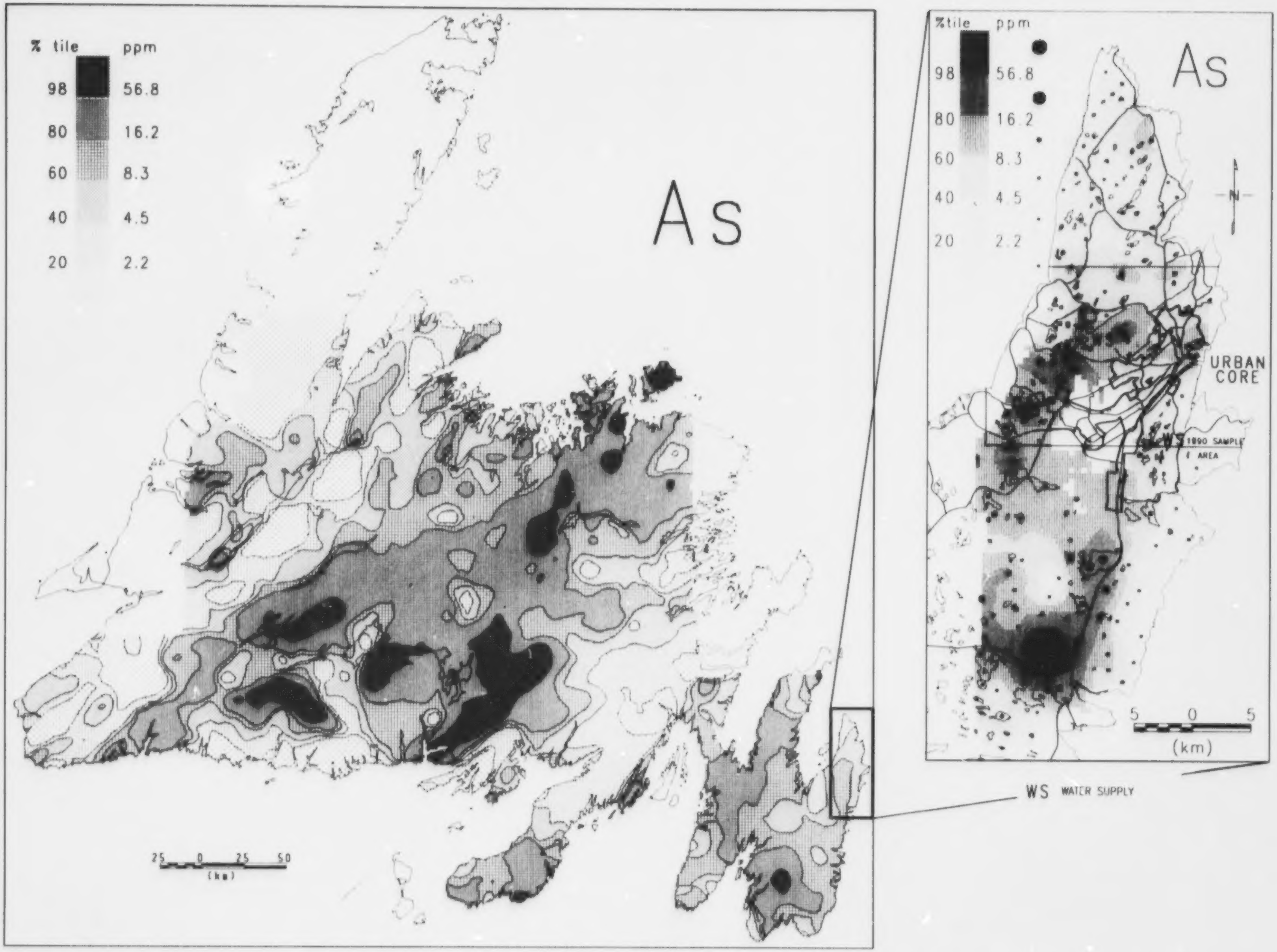
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80
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60
8.3
40
4.5
20
2.2
As
N
URBAN
CORE
WS
1990 SAMPLE AREA
5
0
5
(km)
WS WATER SUPPLY

Figure 9. *Regional distribution of As in lake sediment throughout Newfoundland (left) and, in more detail on the right, from 246 sites on the northeast Avalon Peninsula, including the St. John's urban region.*

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THE DEVELOPMENT AND APPLICATION OF GEOCHEMICAL ANALYSES OF WATER

C. Finch, G. Hall[1] and J. McConnell
Geochemistry and Geophysics Section

ABSTRACT

Methods of sampling, analyzing and interpreting the results of water samples from streams and lakes in Newfoundland are being developed jointly by the Geological Survey Branch and the Geological Survey of Canada. Water samples were collected from 176 sites in 1991 and have been analyzed for over 50 elements by several methods; many elements have been analyzed by two or more methods permitting determination of preferred techniques. Water data will be compared with stream sediment, overbank and lake-sediment data as appropriate. Preliminary work indicates that some water data (e.g., arsenic) give distribution patterns similar to those of lake-sediment data. Water data promise to be useful in the search for mineral resources and for assessing environmental parameters.

INTRODUCTION

A scheme for the rapid analysis of natural fresh water samples for a wide range of elements to levels below their average abundances is being developed based on the techniques of inductively coupled plasma emission spectrometry (ICP-ES), inductively coupled plasma mass spectrometry (ICP-MS), and atomic absorption spectrophotometry (AA). The methods developed under this project, which is being conducted jointly by the Geological Survey Branch and the Geological Survey of Canada, will be used to evaluate the effectiveness of water as a sample medium for geochemical exploration. The project brings together results and techniques from several earlier projects both in Newfoundland (Christopher *et al.*, 1991) and elsewhere in Canada, and involves laboratories at the Geological Survey of Canada (AA and ICP-MS), Memorial University of Newfoundland (ICP-MS) and the Geological Survey Branch (ICP-ES).

SAMPLING AND SAMPLE PREPARATION PROCEDURES

In June 1991, a water-sampling program was conducted in which 40 stream-water samples were obtained from the Baie Verte and Springdale peninsulas and the St. Alban's area, and 136 lake-water samples were collected in the Bay d'Espoir—Fortune Bay area (Figure 1). Seven pairs of site duplicates were collected in the stream survey and 12 pairs in the lake survey. Most of the stream sites had been sampled for overbank and active sediment in 1990 and at the same time vegetation samples were collected (Dunn, 1991), thus the water data may be compared with the data from other media for effectiveness in reflecting mineralization and catchment basin geology. The stream-sediment data have strong geochemical relief in base metals, arsenic, antimony and gold. The areas of lake-water sampling were selected in

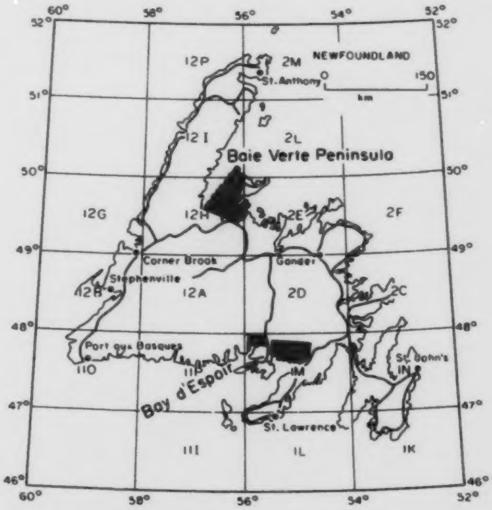


Figure 1. *Location of survey areas.*

This project is funded by the Canada—Newfoundland Cooperation Agreement on Mineral Development (1990-1994); project carried by Geological Survey Branch, Department of Mines and Energy, Government of Newfoundland and Labrador.

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, K1A 0E8

the expectation, based on data from reconnaissance lake-sediment surveys, that the water chemistry would show strong relief in base metals, arsenic, antimony, and rare-earth elements.

Stream and lake water were collected by immersing a 250 ml nalgene bottle beneath the surface. Lakes were sampled in their centres using a float-equipped helicopter. Streams were also sampled near their channel-centres to obtain a homogeneous sample and to avoid undue contributions from local groundwater sources.

Within 24 hours of collection and following analysis for pH and total dissolved solids, the samples were filtered through a 0.45 μ-m filter paper and acidified with 2 ml of nano-pure HNO_3 to await further analyses.

ANALYTICAL METHODS

Table 1 summarizes the suites of elements determined thus far in the three laboratories and the methods used. The overlap in the suites will facilitate a comparison of analytical methods. Table 2 summarizes the work to be accomplished this winter (1991-1992).

The methods employed for the analysis of these waters are Direct Inductively Coupled Plasma Mass Spectrometry (ICP-MS), Hydride Generation Inductively Coupled Plasma Mass Spectrometry (HG-ICP-MS), Ion Chromatography Inductively Coupled Plasma Mass Spectrometry (IC-ICP-MS), Inductively Coupled Plasma Emission Spectrometry (ICP-ES), Ultrasonic Nebulization Inductively Coupled Plasma Emission Spectrometry (USN-ICP-ES), Direct Flame Atomic Absorption Spectrometry (F-AA), Hydride Generation Atomic Absorption Spectrometry (HG-AA), Graphite Furnace Atomic Absorption Spectrometry (GF-AA) and Ion Selective Electrode (ISE).

Samples analyzed by direct ICP-MS underwent no additional sample preparation or preconcentration. Thirty-nine elements were measured on the acidified samples by conventional nebulization techniques (Table 1). An internal standard consisting of Sc, Y, Th, and Tb was pumped with a peristaltic pump and mixed with the sample solution before nebulization and measured by a Siex Elan ICP-MS.

The F-AA determination of Zn, Fe, and Mn was performed directly on the acidified water samples with a Perkin Elmer 3030 Spectrometer. Sodium, K, Ca and Mg were determined by AA, but the samples were made up to 0.1 percent Cs as an ionization buffer and 0.1 percent La as a releasing agent. A 10 ml portion of unacidified sample was buffered with TISAB III after which fluoride was measured by ISE.

Hydride generation methods required sample pretreatment to obtain the desired oxidation state of the elements in solution. Arsenic was measured by HG-AA. Sample pretreatment for this method involved taking 2.5 ml of sample, 0.5 ml of 12 M HCl and 50 micro-litres of a mixture of 2.5 percent KI and 2.5 percent ascorbic acid. The samples were left for several hours to allow reduction of As (V) to As (III). The volatile hydride AsH3 (arsine) was formed with the addition of $NaBH_4$ in a flow-injection system wherein the gases evolved were swept into the a heated quartz tube that was placed in the light-path of a Perkin-Elmer 5000 AA upon which As was measured.

Bismuth, Sb, Se and Te were determined by HG-ICP-MS. An aliquot of sample was pre-reduced overnight with the addition of 0.01 percent KI to reduce Bi (V) and Sb (V) to (III). A second aliquot of sample solution was boiled for thirty minutes in 4M HCl to reduce Se (VI) and Te (VI) to (IV). These elements were then measured by HG-ICP-MS at masses ^{78}Se, ^{121}Sb, ^{130}Te and ^{209}Bi using a GSC-designed hydride generation system and a Perkin-Elmer/Siex Elan 250 ICP-MS.

Rare-earth elements (REEs) are being analyzed by IC-ICP-MS at the GSC. The REEs are present in natural waters at concentrations below detection limits of even the mostly highly sensitive techniques. Research at the GSC has resulted in the ability to measure these elements in the ppt (ng/L) range using ICP-MS after 10-fold preconcentration. A chelating column, the MetPac CC-1 from Dionex Corporation, serves to concentrate and separate the REEs from the alkali and alkaline-earth elements prior to analysis and remove these possible interferences. Although this step requires about 25 minutes, it is completely automated and a batch of samples can be processed overnight without operator attendance. The resin employed is a highly cross-linked, macroporous polystyrene-divinylbenzene containing the iminodiacetate functional group, similar to Chelex-100 but without the undesirable properties of shrinking or swelling. The sample is buffered to a pH of 5.5 and a 50 ml volume pumped through the column. After washing with ammonium acetate, the REEs are eluted with 2M HNO_3 to a final volume of 5.0 ml. This solution is then nebulized into the ICP-MS for determination of the REEs. Corrections are made for the spectral interferences of the oxides of the light REEs on the isotopes of the heavy REEs. Recoveries for all the REEs are about 95 ± 2 percent and calibration is performed using solutions carried through the preconcentration.

Separate unacidified one-litre samples were collected in polyethylene bottles for Au analysis (selected sites). Any Au adsorbed on to the walls of the containers was leached off by the addition of 10 ml of 5 percent Br_2-HCl per litre of sample. After 48 hours of contact time, the sample was transferred to a beaker and evaporated to dryness. The Au was redissolved in 7 ml of 0.5 percent Br_2-HBr. Seven ml of water were added and the Au was extracted into MIBK (methyl isobutyl ketone). Any Fe present was back extracted into 0.1M HBr and the solvent layer analyzed by graphite furnace AA, as described in Hall *et al.* (1986).

The samples have been analyzed by Inductively Coupled Plasma Emission Spectrometry (ICP-ES) using an ARL 3520 Sequential Spectrometer for the elements listed in Table 1.

Table 1. Analyses performed to date by participating laboratory. (NDME–Newfoundland Department of Mines and Energy; USN–ultrasonic nebulizer; MUN/CERR–Memorial University / Centre for Earth Resources Research; GSC–Geological Survey of Canada; HG–hydride generation; ISE–ion-selective electrode; HG-AA–hydride generation atomic absorption spectrophotometry; F–flame)

NDME Laboratory			MUN/CERR		GSC Laboratory				
ICP-ES	USN-ICP-ES		ICP-MS		HG-ICP-MS	ISE	HG-AA	F-AA	GF-AA
Si	Li	Cr	Li	Br	Se	F	As	Na	Au
Ca	Be	Mn	Be	Se	Te			K	
Mg	B	Co	B	Rb	Bi			Ca	
Na	Al	Ni	Mg	Sr	Sb			Mg	
K	P	Cu	Al	Mo				Mn	
S	Fe	Zn	Si	Ag				Fe	
Fe	Sr	Y	P	Cd				Zn	
Mn	Ba	Mo	S	Sn					
	Ti		Cl	Sb					
			Ca	I					
			Ti	Cs					
			V	Ba					
			Cr	La					
			Mn	Ce					
			Fe	Hg					
			Co	Tl					
			Ni	Pb					
			Cu	Bi					
			Zn	U					
			As						

Table 2. Analyses to be performed during winter of 1991-92

NDME Laboratory		GSC Laboratory		
ICP-ES	ICP-hydride generation	IC-ICP-MS	IC	UV/Conductivity
C	As, Se, Sn, Sb, Te Hg, Pb, Bi	REEs (i.e., La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu)	SO_4^{2-}	Organic C
		Y	PO_4^{3-}	
		Mn	Cl^-	
		Co	Br^-	
		Ni	NO_3^-	
		Cu		
		Zn		
		Cd		
		In		
		Pb		

IC–ion chromatography; UV–ultraviolet

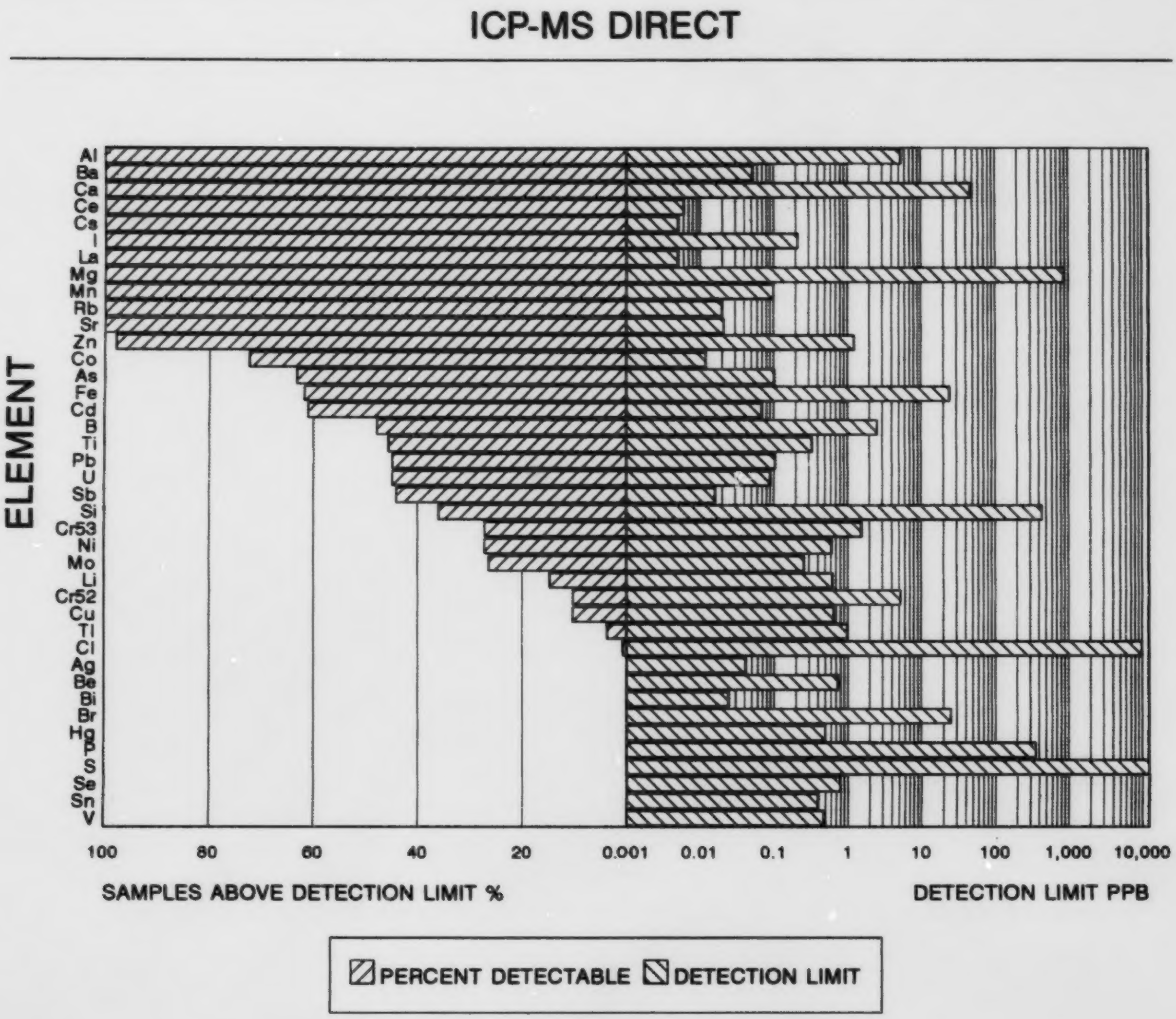


Figure 2. *Summary of percentages of samples with detectable concentrations and detection limits for ICP-MS direct analyses.*

Determination of elements by Ultrasonic Nebulization Inductively Coupled Plasma Emission Spectrometry (USN-ICP-ES) and by Hydride Generation Inductively Coupled Plasma Emission Spectrometry (HG-ICP-ES) are being analyzed at the time of writing at the Geological Survey Branch Laboratory.

The ICP-ES and USN-ICP-ES do not require any sample pretreatment, but USN-ICP-ES provides a five- to ten-fold improvement in detection limits for most elements compared with direct ICP-ES. This improvement is accomplished by desolvation of the sample matrix before introduction to the ICP plasma, which effectively provides a preconcentration of the sample. The development work for HG-ICP-ES methods will be carried out this winter (1991-1992) at the Geological Survey Branch Laboratory. The major objective will be to provide a suite of elements that can be routinely determined in water samples at the Geological Survey Branch, Department of Mines and Energy, for applications in exploration and environmental geochemistry.

RESULTS

For the elements analyzed, there is a wide range in the observed detection limits and the percentages of samples having detectable concentrations. Detection limits range from 0.005 ppb (cesium) by ICP-MS to 10,000 ppb (sulphur), also by ICP-MS. The detection limit for S is much lower, however, by ICP-ES. Several elements have no detectable concentrations in any of the samples whereas 13 elements were detected in all samples. These observations are quantified graphically in Figures 2 and 3.

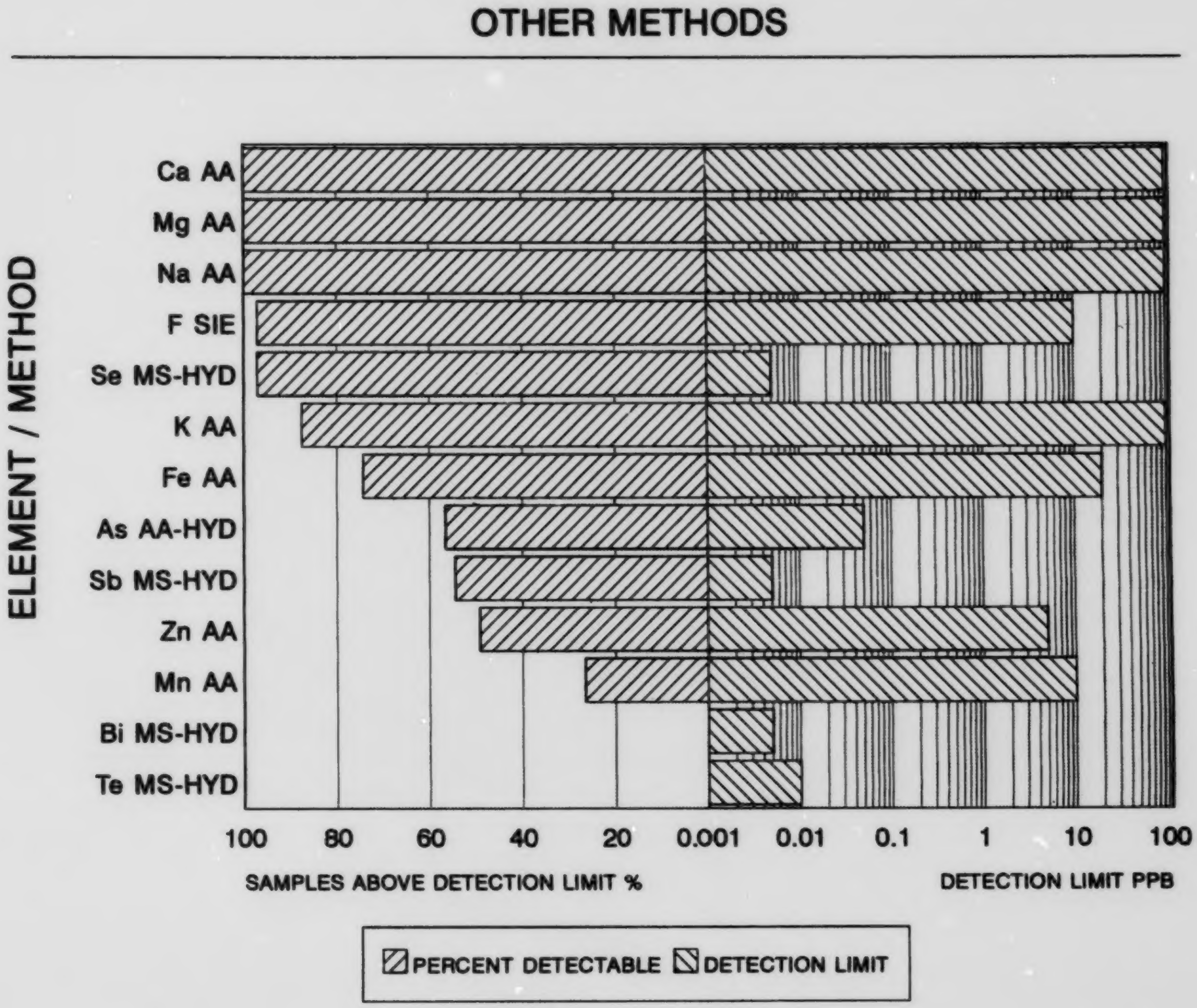


Figure 3. *Summary of percentages of samples with detectable concentrations and detection limits for analytical methods other than ICP-MS direct analyses.*

Comparison of Methods

Several elements were analyzed by two methods. The detection limits and percentage of samples analyzed with detectable concentrations for nine elements are plotted in Figure 4. Results of two methods of analysis can also be compared by reference to Figures 5 and 6, in which data from 18 pairs of site duplicates are shown in scatterplots. For Ca and Mg, the two methods, AA and ICP-MS, give excellent and comparable results. All samples have detectable concentrations and the correlation coefficients of 18 pairs of site duplicates exceed 0.99 (Figure 6).

Analyses of iron (Fe) by AA appear preferable to the results by ICP-MS (Figure 4). Not only do more samples have detectable Fe in the AA method but the correlation coefficient of site-duplicate pairs by AA is 0.99 compared to 0.48 by ICP-MS (Figure 5).

Both methods of analysis (ICP-MS and AA-hydride) for arsenic (As) give good and comparable results. The ICP-MS data seem slightly preferable as the correlation coefficient for site-duplicate pairs is stronger, 0.998 compared with 0.989 for AA-hydride; however, the hydride method has a marginally lower detection limit.

Analysis of antimony (Sb) is better by MS-hydride than by ICP-MS. The lower detection limit by the hydride method gives considerably more samples with detectable concentrations. The correlation coefficient for Sb in the site duplicates is lower by the hydride method (0.77) than by ICP-MS (0.94) due to the inclusion of a single 'flyer' in the data.

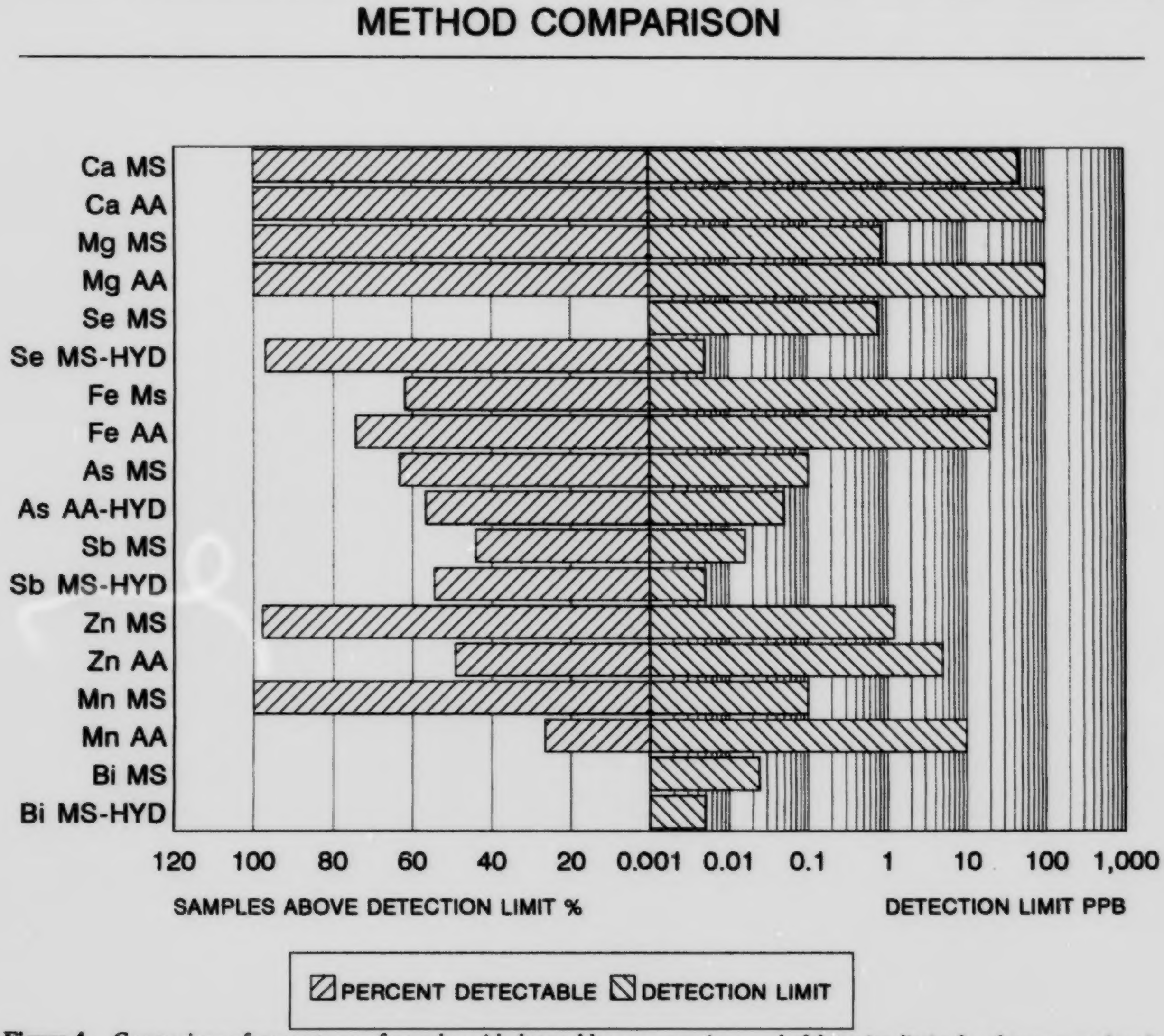


Figure 4. *Comparison of percentages of samples with detectable concentrations and of detection limits for elements analyzed by two methods.*

If this is removed, the remaining pairs have a correlation coefficient of 0.98.

The results for zinc (Zn) by ICP-MS and by AA differ markedly and neither dataset is satisfactory. Although the ICP-MS method apparently offers a lower detection limit and has more samples with detectable Zn, there is no correlation between analyses of site-duplicate pairs. The correlation coefficient for the ICP-MS data is -0.02 whereas for the AA data, it is 0.62. However, the stronger coefficient for the AA data is due almost entirely to one sample pair with high Zn in both analyses.

Manganese (Mn) was analyzed by ICP-MS and by F-AA. The former method gives much more useful data because of its significantly lower detection limit. The reproducibility of analyses is very good by both methods with correlation coefficients of 0.92 and 0.99 for site-duplicate pairs by ICP-MS and F-AA, respectively.

Bismuth (Bi) has no detectable concentration in these samples by either ICP-MS or MS-hydride. For selenium (Se), only the MS-hydride method gives useful data.

Reproducibility of data-analysis of site-duplicate pairs

One method of assessing the significance and variance of the analytical data is to evaluate the results of analyses of site-duplicate samples. This is done here by presenting scatterplots and Pearson correlation coefficients of data from 18 pairs of site duplicates (Figures 5 and 6). Elements included are those for which more than 40 percent of the samples analyzed by the three major methods have detectable

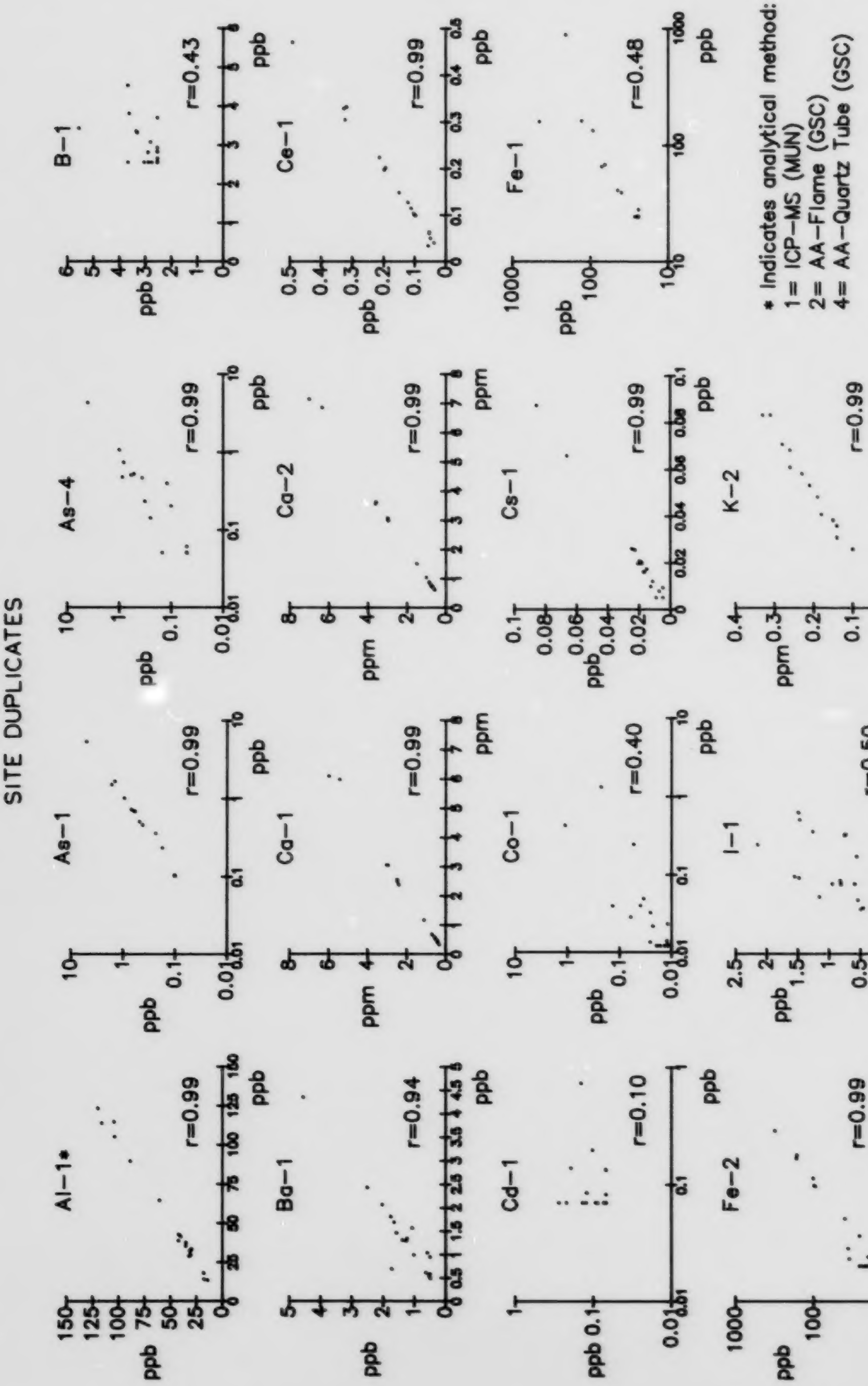


Figure 5. *Scatterplots of analyses of site duplicates for aluminium, arsenic, boron, barium, calcium, cerium, cadmium, cobalt, cesium, iron, iodine, and potassium. (Note: log axes for As, Cd, Co and Fe).*

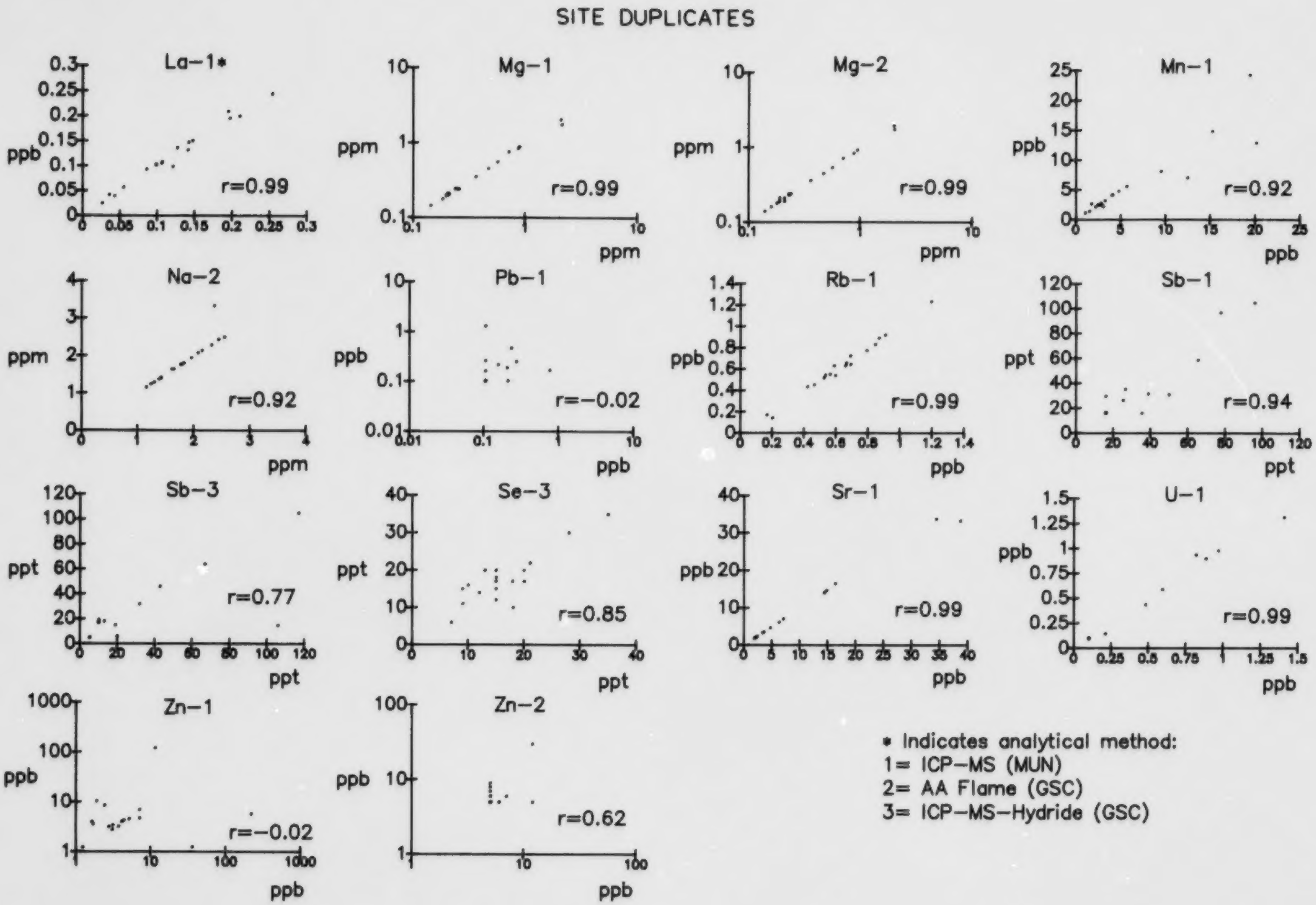


Figure 6. *Scatterplots of analyses of site duplicates for lanthanum, magnesium, manganese, sodium, lead, antimony, selenium, strontium, uranium and zinc. (Note: log axes for Mg, Pb and Zn).*

concentrations. For ICP-MS, these elements are: Al, As, B, Ba, Ca, Ce, Cd, Co, Cs, Fe, I, La, Mg, Mn, Pb, Rb, Sb, Sr, U and Zn. For AA, the elements include: Ca, Fe, K, Mg, Na and Zn. For MS-hydride the elements are Sb and Se and for AA-hydride the element is As.

The results of individual elements analyzed by two separate methods have been discussed above. Of the remaining elements, most analyses reproduce with good to excellent correlations. These include Al, Ba, Ce, Cs, K, La, Mn, Na, Rb, Se, Sr and U. Fair correlation is seen in B, Co and I. Correlation is poor for Cd.

The REE data are given in Table 3 for five separate analyses of two control samples (waters from the Ottawa River and Meech Lake). Relative standard deviations are generally better than 10 percent, even at levels of REEs below 10 ppt. This preconcentration procedure is also effective for the first-row transition elements, Cd and Pb (Siriraks *et al.*, 1990). These results when completed will be compared to those by direct ICP-MS.

Table 3. Reproducibilities of REE analysis by IC-ICP-MS

Element	Ottawa River (n=5) mean,ppt	Ottawa River (n=5) RSD,%*	Meech Lake (n=5) mean,ppt	Meech Lake (n=5) RSD,%
Y	141	2.2	42	2.8
La	270	2.5	33	2.3
Ce	335	1.6	42	4.0
Pr	63	2.7	7	2.4
Nd	251	3.5	29	8.4
Sm	41	5.3	5	12
Eu	9	7.3	7	4.5
Gd	31	15	5	15
Tb	4	5.7	0.7	21
Dy	20	4.5	4	10
Ho	4	3.7	1	9.2
Er	11	5.3	3	6.6
Tm	0.5	10	0.5	9.2
Yb	11	4.9	4	8.2
Lu	2	13	0.8	17

* Relative Standard Deviation

Chondrite-normalized plots (Figure 7) are shown for two stream waters and two lake waters from the Bay d'Espoir area. Note the distinct, negative, Ce anomalies for the stream waters, possibly indicating input from marine aerosols. Both lakes show sharp negative Eu anomalies. The corresponding sediment plots (Figure 8)—derived from neutron-activation analysis for La, Ce, Sm, Eu, Tb, and Yb—also feature Eu anomalies. The one for stream sediment 91-123 is much more pronounced than is observed for its water.

APPLICATIONS

One method of portraying the distribution of elements in the physical landscape is provided by contoured data maps.

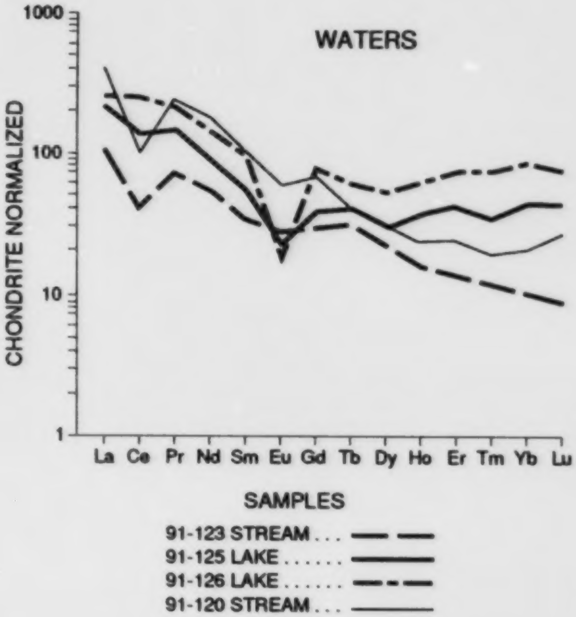


Figure 7. *Chondrite normalized plots of REEs for stream and lake-water samples from the Bay d'Espoir area.*

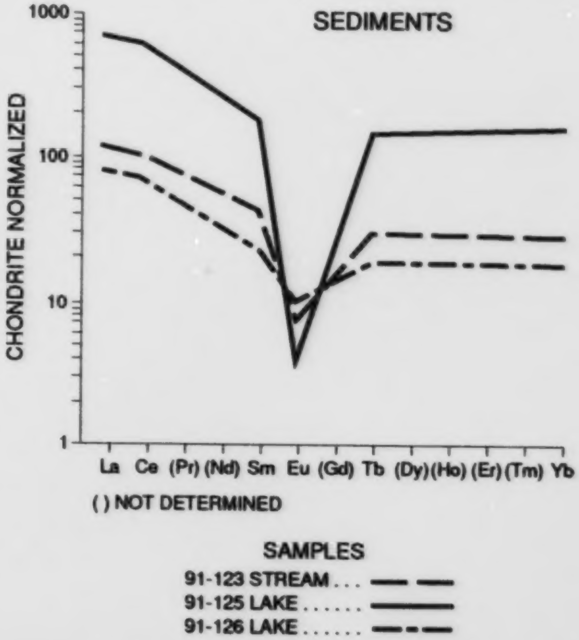


Figure 8. *Chondrite normalized plots of REEs for stream and lake-sediment samples from the Bay d'Espoir area.*

The distributions of arsenic in lake water and arsenic in lake sediment in the Bay d'Espoir area are illustrated in Figures 9 and 10. Sample sites are shown by '+' symbols. The lake-sediment samples were collected in 1978 (Davenport *et al.*, 1990) and were not necessarily collected from the same lakes

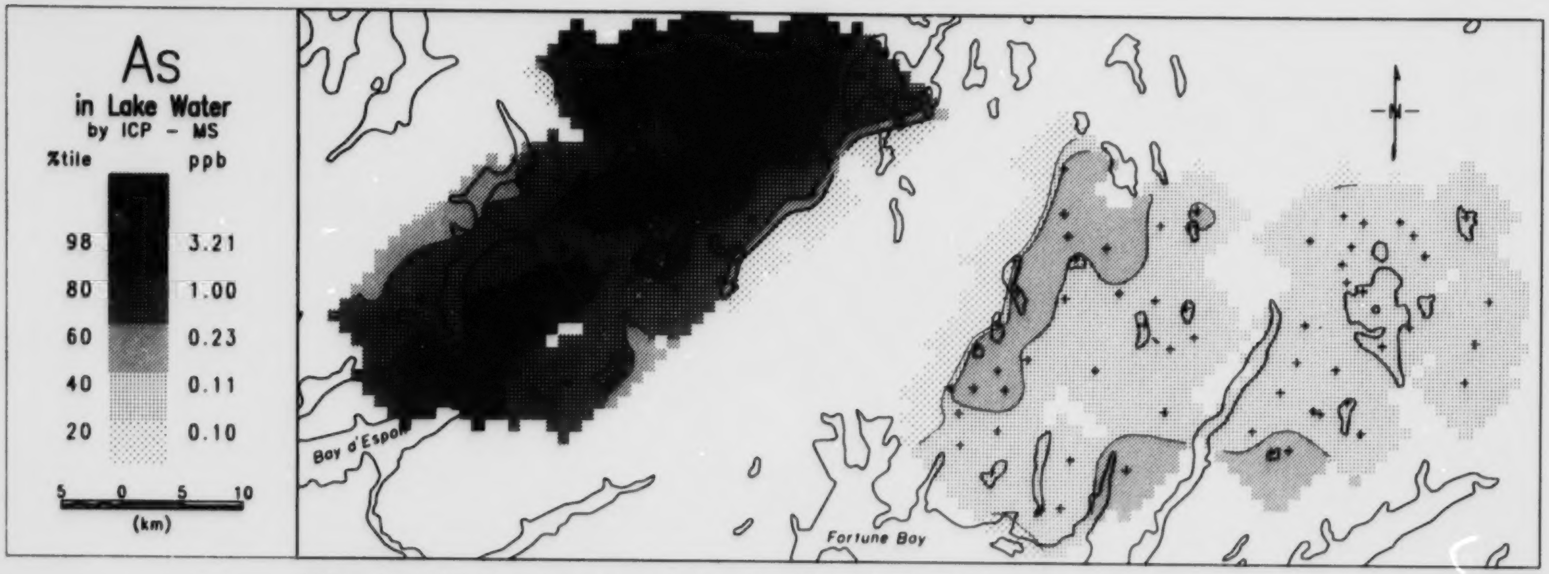
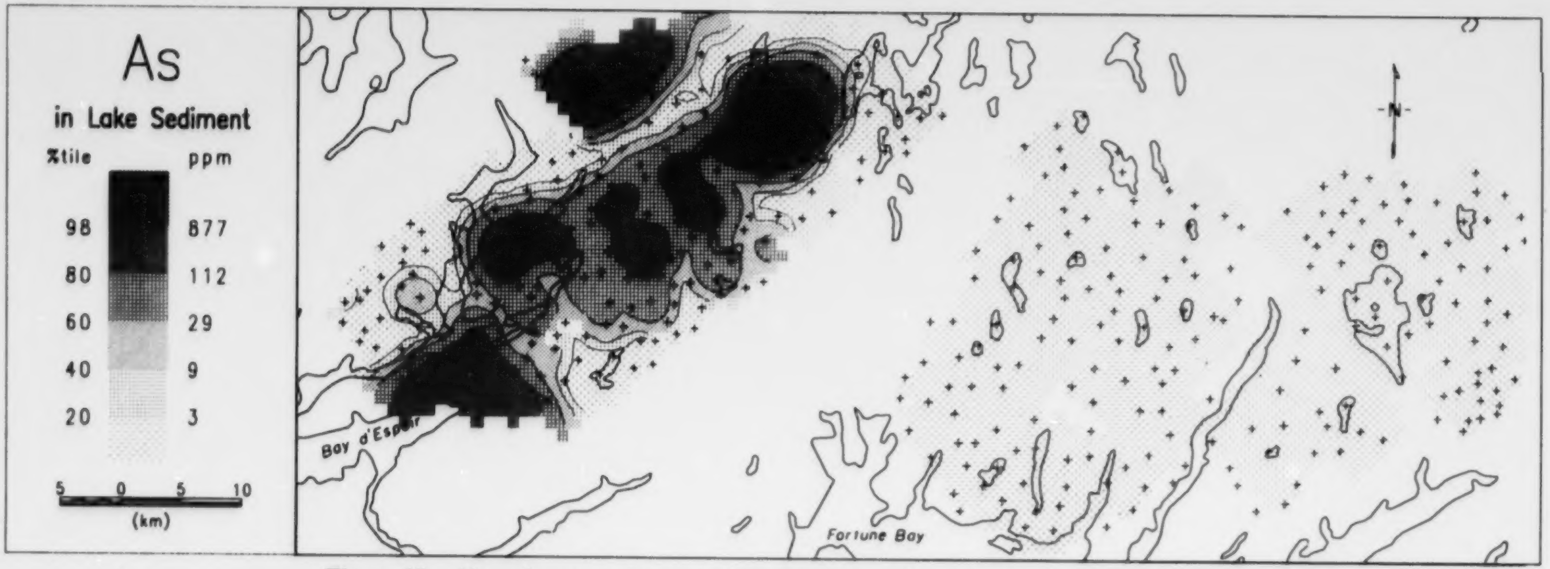

As
in Lake Water
by ICP - MS
%tile
ppb
98
3.21
80
1.00
60
0.23
40
0.11
20
0.10
5
0
5
10
(km)
N
Bay d'Espoir
Fortune Bay

Figure 9. *Map of contoured arsenic data from lake water in the Bay d'Espoir area, Newfoundland.*


As
in Lake Sediment
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ppm
98
877
80
112
60
29
40
9
20
3
5
0
5
10
(km)
N
Bay d'Espoir
Fortune Bay

Figure 10. *Map of contoured arsenic data from lake sediment in the Bay d'Espoir area, Newfoundland.*

as were the water samples. The data in each set have been gridded to a 1 by 1 km grid, a process that results in some smoothing of the data. The area to the west in each figure is underlain primarily by clastic sedimentary and volcanic rocks whereas the area to the east is dominated by granitic rocks. The distribution patterns of As in the two media are seen to be broadly similar despite the samples being, in many instances, from different lakes and despite a 13-year gap between the two surveys. The analyses of As in sediment are uniformly low in the eastern area.

CONCLUSIONS

Recently developed analytical techniques permit the determination of a wide range of elements in samples of surface waters. For many elements, the analytical detection limits are sufficiently low to determine element concentrations in most samples. Preliminary comparison of lake water and lake-sediment data from common lakes indicate that some elements can be correlated between the two media.

ACKNOWLEDGMENTS

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RECLAIMING GEOPHYSICAL DATA FROM MINERAL-ASSESSMENT REPORTS

G.J. Kilfoil and P. Honarvar
Geochemistry and Geophysics Section

ABSTRACT

A project was initiated to extract geophysical information in computer-processable (digital) form from mineral-assessment reports, submitted over the last 40 years. A pilot project was designed to initially extract information from only those files containing airborne geophysical surveys for Labrador. The digitized outlines of these surveys have been merged with corresponding GEOSCAN bibliographic information to create a digital index that can easily be queried on a PC. Two methods of capturing the data, by digitizing and optical scanning, from contour maps taken from the assessment reports were investigated for their accuracy and efficiency. At present, despite the time and resource requirements, digitizing is the superior method of data input due to manual control over data entry and minimal data-storage requirements.

Aeromagnetic data from two detailed industry surveys were assessed by comparison to the existing data from surveys conducted by the Geological Survey of Canada. In these examples, the digitally processed data from detailed industry surveys provide better definition of geological features than the more regional survey data.

INTRODUCTION

The objective of this project is to begin to build a well-organized, computer-processable geophysical database for the province, to support mineral exploration and other government and university geoscientific programs. Most of the digital geophysical data now publicly available are from regional surveys conducted by the Geological Survey of Canada. These are being organized and processed using a set of PC-based programs (Kilfoil and Nolan, *this volume*) to form a regional framework for the database.

A considerably greater amount of geophysical data is contained in assessment reports submitted to the Department of Mines and Energy by mineral-exploration companies. These industry surveys are varied in their type, scale, documentation and, we expect, quality. Furthermore, almost all survey data exist in our files only as paper (or mylar) copies or on microfiche. The main part of this project is concerned with filling in the regional framework of digital geophysical data with these industry data in digital form. To accomplish this, the industry surveys must be first catalogued, their quality and level of documentation assessed, and selected surveys digitized, processed and included. To keep the project to a manageable size at the outset, the first surveys to be examined are airborne geophysical surveys reported in non-confidential assessment files for Labrador. This pilot project will allow the design and costing to be determined for a more comprehensive project that would include the whole province and all geophysical surveys, as well as an evaluation of the benefits of such a project.

CONSTRUCTING A DIGITAL INDEX

Catalogues of geophysical and geochemical surveys for the province have been compiled in booklet form from time to time (Mullins *et al.*, 1969; Withycombe and Fogwill, 1978), or as maps (Harris, 1977). For this project, a digital index is being designed. This contains relevant synoptic information from assessment reports linked to digital survey boundaries through a computer database with graphical output. The digital index will permit much more effective search, browse, retrieval and update of the file, and will be eventually linked to the actual data in the geophysical database, probably through a WINDOWS (Microsoft, 1990) application.

By the end of August, 1991, the outlines of all airborne geophysical surveys for Labrador had been digitized from maps submitted as part of the reports in Geofiles. As each survey boundary was input, the survey was classed into three broad categories by the type of survey flown: magnetic, radiometric and electromagnetic. No distinctions were made between the various types of electromagnetic receivers, whether active or passive, used during the survey. Figure 1 shows the outlines of industry-sponsored, airborne geophysical surveys (approximately 120 surveys from 70 assessment reports) flown to date in Labrador, coded by the possible combinations of survey type as seven line types and weights. This coding readily illustrates the distribution of surveys that exist for an area of interest and identifies each by survey type. Note the concentration of surveys in the southern Labrador Trough and over the Central Mineral Belt.

This project is partially funded by the Canada–Newfoundland Cooperation Agreement on Mineral Development (1990-1994); project carried by Geological Survey Branch, Department of Mines and Energy, Government of Newfoundland and Labrador.

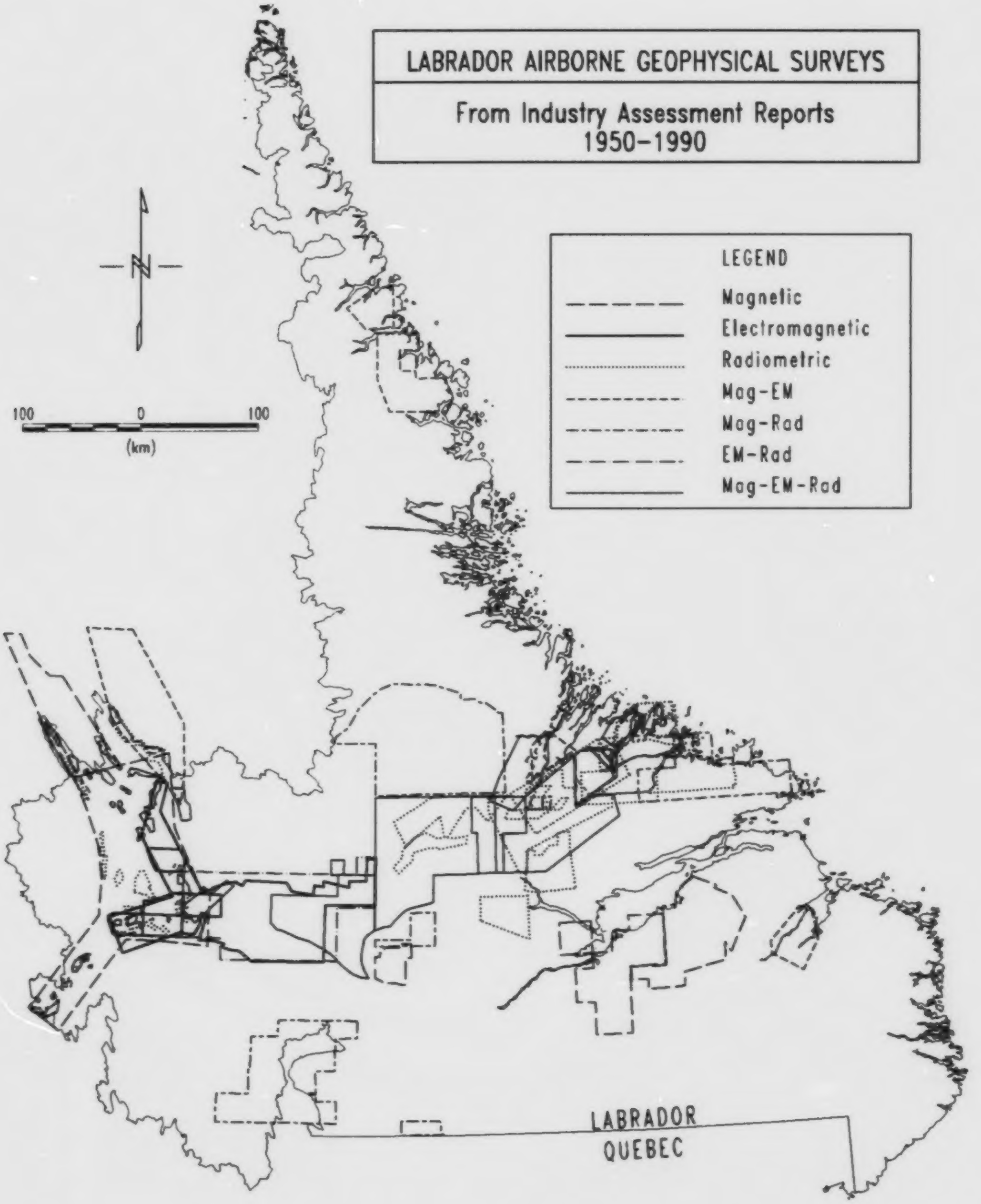
LABRADOR AIRBORNE GEOPHYSICAL SURVEYS
From Industry Assessment Reports
1950–1990
LEGEND
Magnetic
Electromagnetic
Radiometric
Mag-EM
Mag-Rad
EM-Rad
Mag-EM-Rad
N
100
0
100
(km)
LABRADOR
QUEBEC

Figure 1. *Index map of airborne geophysical surveys conducted in Labrador between 1950 and 1990. Surveys are coded by line type according to geophysical method. The map was output from a digital map index database, which consists of survey outlines having corresponding bibliographic information from the GEOSCAN database.*

To form the foundation for a digital index, a relational database was constructed from bibliographic and spatially referenced information. First, selected descriptive text fields were exported from the records of all airborne geophysical surveys conducted in Labrador and contained in the national GEOSCAN bibliographic database (Patey and Gillespie, *this volume*). These were reformatted and imported into a dBase III format (Ashton-Tate, 1988) database file. The UTM coordinates defining the individual survey outlines were digitized, reformatted and written to a memo file and referenced to a separate relational database file using the MKPOLY utility from the QUIKMap (Environmental Sciences Limited, 1990) graphical mapping package. Finally, corresponding records from the two relational database files were merged (using a Clipper routine) to create a digital index database file.

As part of the pilot project, the InFOcus (Earth and Ocean Research, 1990) desktop-mapping software package was evaluated for its suitability as a graphical interface to the digital index database. The InFOcus package is a menu-driven link to information contained in a relational database and also provides a user-friendly selection of QUIKMap utilities. In turn, QUIKMap provides the interface from the database to output devices, such as a VGA monitor or line plotters. The QUIKMap utilities provide a mouse-driven graphical display that enables convenient zooming and point-and-click selection of polygonal survey outlines to access data associated with survey areas. An option in the InFOcus menu allows vector information existing in other databases, such as the outlines of claim blocks, geological contacts or topographic features, to be selected for overlay as base maps, on the graphical image of survey outlines. Written in FoxPro (Fox Software Inc., 1990), InFOcus quickly accesses information concerning a selected survey in the database file and presents it in an adjustable text box that overlays on top of the graphical image. Information displayed in the text box can be interactively edited. A multitude of criterion-based searches can be instituted from the InFOcus menus; records meeting the search criteria are quickly flagged for display through the efficient database access of FoxPro. The combination of InFOcus coupled to QUIKMap also provides the ability to select data for output to plotters at any specified scale or projection.

Although the InFOcus software provides efficient database access, display, editing and zooming capabilities, a few inherent shortcomings limit its suitability as a graphical interface for the digital index. First, the procedure required to construct the digital index database is not intuitive. As well, a single assessment report commonly describes geophysical surveys carried out in more than one area. Conversely, more than one assessment report may be submitted for a single survey. No simple method is apparent for establishing such complex links between survey outlines and textural information within the database. If the areas enclosed by survey outlines overlap and the screen pointer is clicked in the overlap area, only information from the first record, which corresponds to one of the areas encountered in the database, will be displayed in the text box. In short, a simultaneous query of information from multiple database records that may exist for a specified location cannot be easily accomplished within the current version of InFOcus.

Alternative software packages upon which to base the digital map index are currently being sought. Its functional and adaptable capabilities, ease of use and cost are seen as the major criteria in graphical interface selection. The graphical database access software should be reasonably priced to ensure the widespread use of the digital index by mineral-exploration companies. The compiled index and digitized outlines of airborne geophysical surveys for Labrador currently exists as a relational database in dBase III format and as ASCII text files, either of which could be reformatted for import to an alternative database structure.

A digital index has several advantages over the traditional booklet or index map. Being PC-based, bibliographic information can be retrieved faster and more efficiently. The digital index is more comprehensive in that all existing assessment-report information can be combined into a single database, which can be readily updated, and from which the user can select the level of bibliographic information to be output or displayed on the computer screen. The graphically oriented display of survey boundaries is more visually appealing to the user and, through zooming facilities, the level of spatial detail to be displayed, or output, is unconstrained. When fully developed and linked to geophysical data files from individual surveys, the digital index will theoretically be the 'roadmap' for navigating through the geophysical database.

DIGITAL DATA INPUT (DIGITIZING VERSUS SCANNING)

Two methods of capturing geophysical data from analog paper maps or plots were evaluated and the merits of each compared: 1) selective input from a digitizer, or 2) total input by digital scanner, followed by selective editing of the converted vector data.

As an example, a map from an assessment report covering the 1982 survey flown in the Florence Lake area, Labrador (BP Minerals Ltd. and Billiton Canada Ltd., 1983) was selected to compare the two data-capture methods. In particular, the contoured, total-field aeromagnetic map for the northern part of the Florence Lake survey was selected because the lines comprising this computer-generated map are relatively uniform in weight and density, the map area is of manageable size for scanning, and a mylar copy (not as prone to paper shrinkage or distortion) was available.

Since the data on the map consists of contours and flight-line paths with fiducial points, it is impossible to completely recover the original raw survey data. On maps such as these, the intersections of flight lines with contours are the points where the original data are considered to be best represented (Kilfoil, 1990) because only along flight lines are data collected that will constrain the position of contours. Therefore, the aim in capturing the data digitally is to obtain a sequence of X and Y coordinates with associated data values at contour intersections along the flight lines.

The map was converted to a raster image by a large format scanner with 400 dots/inch resolution. The raster image was then vectorized into line segments, in AutoCAD format, by GTX conversion hardware/software. The raster image, stored in a binary (PCX) format, requires approximately 5 megabytes of disk storage space; the converted file of vector line segments for the map, stored (as

ASCII) in the AutoCAD Data eXchange Format (DXF), occupies 7.7 megabytes of disk storage space. The X and Y coordinates of points comprising vector line segments are in thousandths of an inch at the map scale. The vector line segments in the AutoCAD DXF file were plotted on a pen plotter (see Figure 2). The darker areas on Figure 2 indicate locations where adjacent lines, which are close together and/or thick on the original map, have merged on output to a single line or a filled area. These effects are caused by a combination of the finite scanner resolution and locally degraded quality on the original map. As well, contour lines are incomplete or discontinuous in areas (on Figure 2) where the original map lines are faint. This results from the inability of the digital scanner to adjust to non-uniform line quality on the original.

Each contour line segment in the vectorized data-file includes positional information along its total length, whereas only along flight lines is this information constrained by original survey data. Scanned contour line segments do not have an associated attribute to reflect the magnetic field strength values that they represent. The vector file also contains a substantial proportion of extraneous information (map borders, titles, north arrow on Figure 2) that, in this vector format, are not essential to the geophysical content of the map. Thus, in order to obtain a datafile that approximates the original geophysical data, the scanned digital data must undergo extensive editing, followed by a conversion of map coordinates to a geographic coordinate system. Due to the large size of the vector datafile that results from scanning this map, such editing has not yet been attempted, as the present editing software limitations would significantly slow extraction of relevant information.

The intersections of all contours with flight lines were also digitized manually from the same aeromagnetic map that was scanned. The digital data were subsequently edited, gridded and processed to remove any inter-profile levelling errors. The data were then contoured (Figure 3) for comparison with the map from which the data were digitized.

A major advantage of digitizing over scanning is that the operator is selective and systematic about data entry. Datafiles generated from digitizing are of significantly smaller size (two orders of magnitude) than the equivalent scanned images and, if points are digitized in sequence along each flight line, are more logically organized for subsequent data processing. The digitized aeromagnetic data for the northern part of the Florence Lake survey required only 70 kilobytes (in ASCII) of disk-storage space. The quality of datafiles that result from digitizing is not nearly as dependent on original map quality as that which results from optical scanning.

Among the obvious disadvantages of digitizing data from a contoured map are the slow progress and tediousness that are inherent in the task. For instance, the digitizing of the data from which Figure 3 was produced required three days to complete, and to digitize all the geophysical data from this one assessment report would take one person about one month to complete. However, the size of scanned images (and resulting vector files) are too large to be effectively edited at present, and, if efficient software were available for this purpose, the time and resource requirements could very well exceed those for digitizing.

Digitized and/or manually input data is also prone to non-systematic human error, whereas optically scanned data tend to contain only errors of a systematic nature (i.e., errors uniformly distributed throughout the output images). A common example of a non-systematic human error would be incorrect assignment of the data value to a flight-line/contour intersection by the user. This type of error can often be detected and rectified during subsequent digital processing of the data. These errors are not confined to digitized files either, as vector files that result from optical scanning are also subject to human errors if such output requires subsequent interactive editing by an operator.

ASSESSING DATA QUALITY

Having input the aeromagnetic data from a contour map in the form of data profiles, it can be easily checked for errors through an iterative process of alternately gridding and checking data values within profiles. By interactively generating a series of colour, shaded-relief images of the gridded datafile on a PC using the in-house DISPLAY program (Kilfoil and Nolan, *this volume*), even very small errors in the data will become evident as subtle discontinuities. The locations of discontinuities in the gridded data can be checked on the original contour map for the existence of an erroneous value. A graphical profile editing program allows a specific profile to be rapidly located, an erroneous data point to be pin-pointed along the profile, and the data value to be corrected interactively.

After correcting any detected errors in the input data, the quality of the survey needs to be assessed. The most obvious method for assessing data quality is comparison with any existing related dataset for which the data quality is known. A coincidence of anomalies or features in two similar datasets from different surveys helps to establish a confidence level in the data from each survey.

In the case of airborne magnetic data, regional coverage for the entire province has been flown by the Geological Survey of Canada at a nominal flight altitude of 300 m and line spacing of 800 m. In addition, magnetic data were recorded along profiles during surveys of the gamma-ray spectrometric program carried out by the GSC during the 1980's. Currently these surveys, flown at 120-m ground clearance and 1-km flight-line spacing, extend over approximately two-thirds of the Island of Newfoundland.

The data for the area covered by the northern part of the Florence Lake survey was extracted from the levelled, corrected and gridded data from the GSC surveys (Kilfoil, 1991). Since the grid-cell spacing of the extracted data was 200 m, the data was regridded to 50 m to match the grid interval for the BP-Billiton survey. The result was contoured

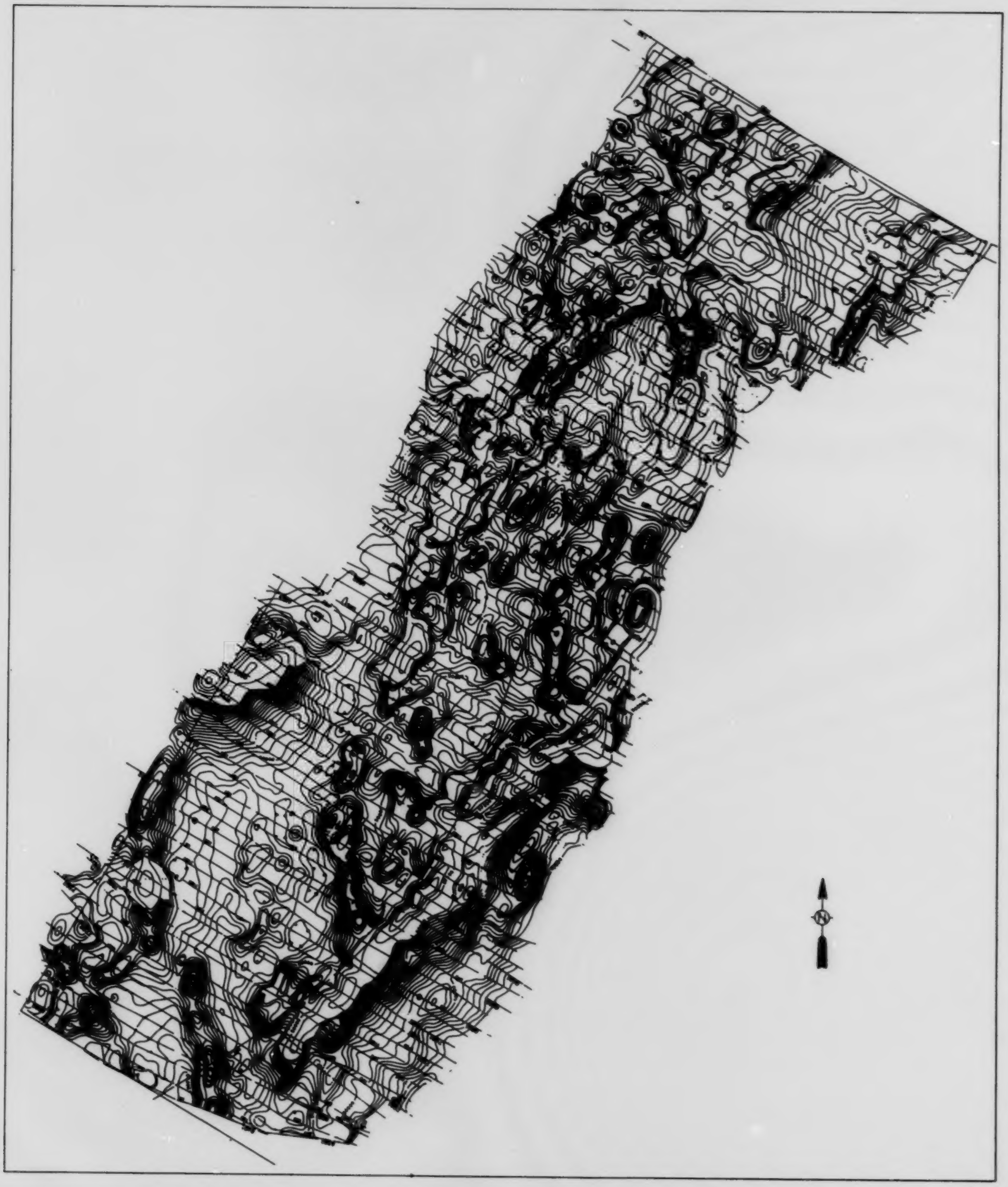

Figure 2. *Plot of an optically scanned, total field aeromagnetic contour map for the Florence Lake (North) area, from a mineral-assessment report submitted by BP-Billiton (1983). The consistently spaced lines on this plot, which strike to the southeast, are the flight-line paths. Certain flight lines and contours appear discontinuous where these lines were faint on the original map. Note also that where lines are closer than the scanning resolution, the subsequent vectorization yields a single thick line or filled area.*

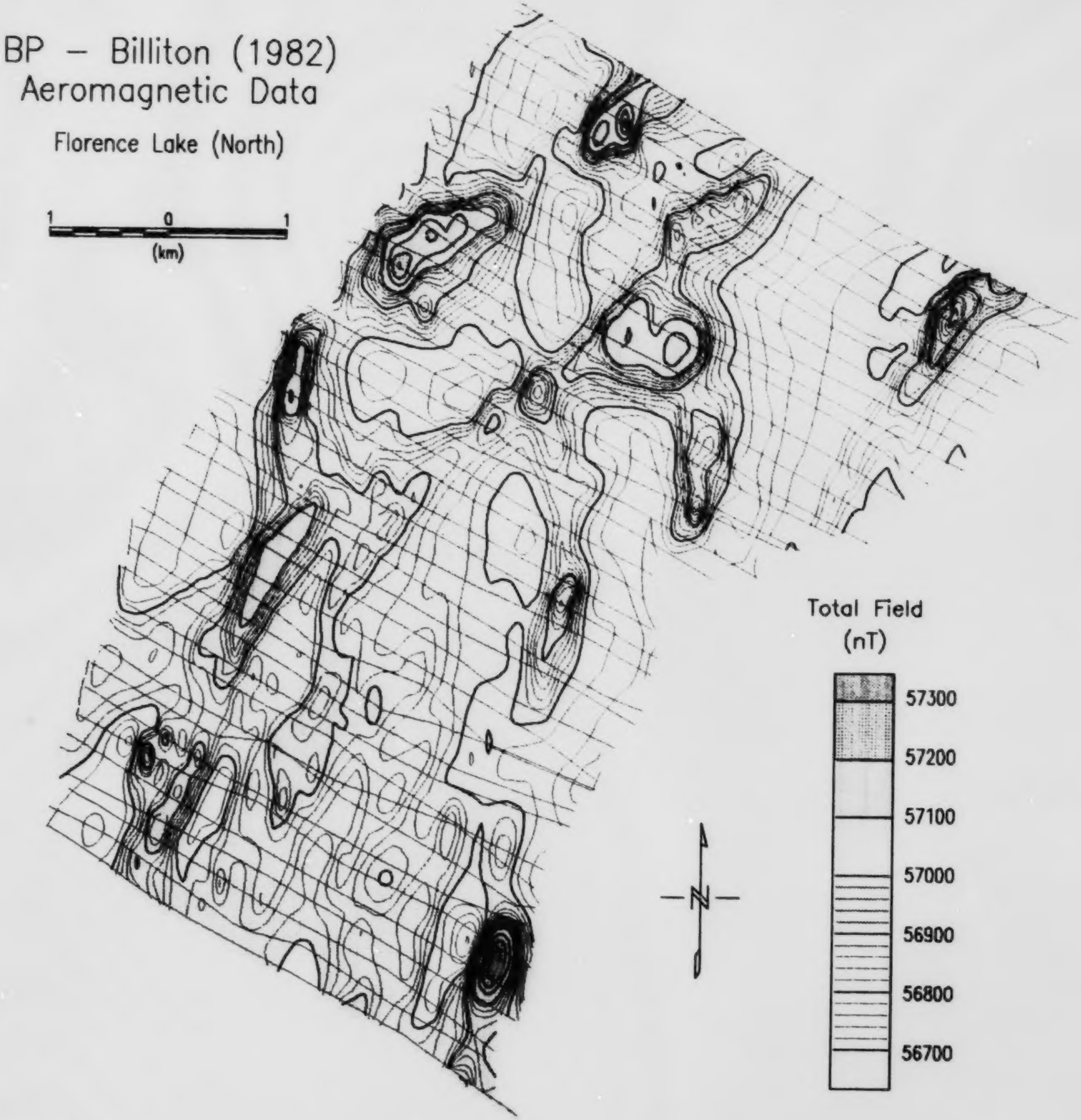
BP – Billiton (1982)
Aeromagnetic Data
Florence Lake (North)
1
0
1
(km)
Total Field
(nT)
57300
57200
57100
57000
56900
56800
56700

Figure 3. *Shaded, aeromagnetic contour map for the Florence Lake (North) area, produced from the digitized location of intersections of contours with flight lines on maps submitted by BP-Billiton in 1983. The area covered is the northeast half of that shown in Figure 2. Total-field values have not been referenced to the International Geomagnetic Reference Field (IGRF), and are variably shaded to indicate anomalous magnetic field intensity. The nominal survey altitude is 30 m and flight-line spacing averages 200 m.*

and is presented here as Figure 4. A quick comparison with Figure 3 reveals a strong correlation of long-wavelength (broad) anomalies, which lends credence to both datasets and indicates that the two surveys are properly geo-referenced. However, the comparatively smooth contours on Figure 4 indicate that the data from the BP-Billiton assessment report, despite being reconstructed by digitizing a contour map, contains significantly more detail. Anomalous zones and

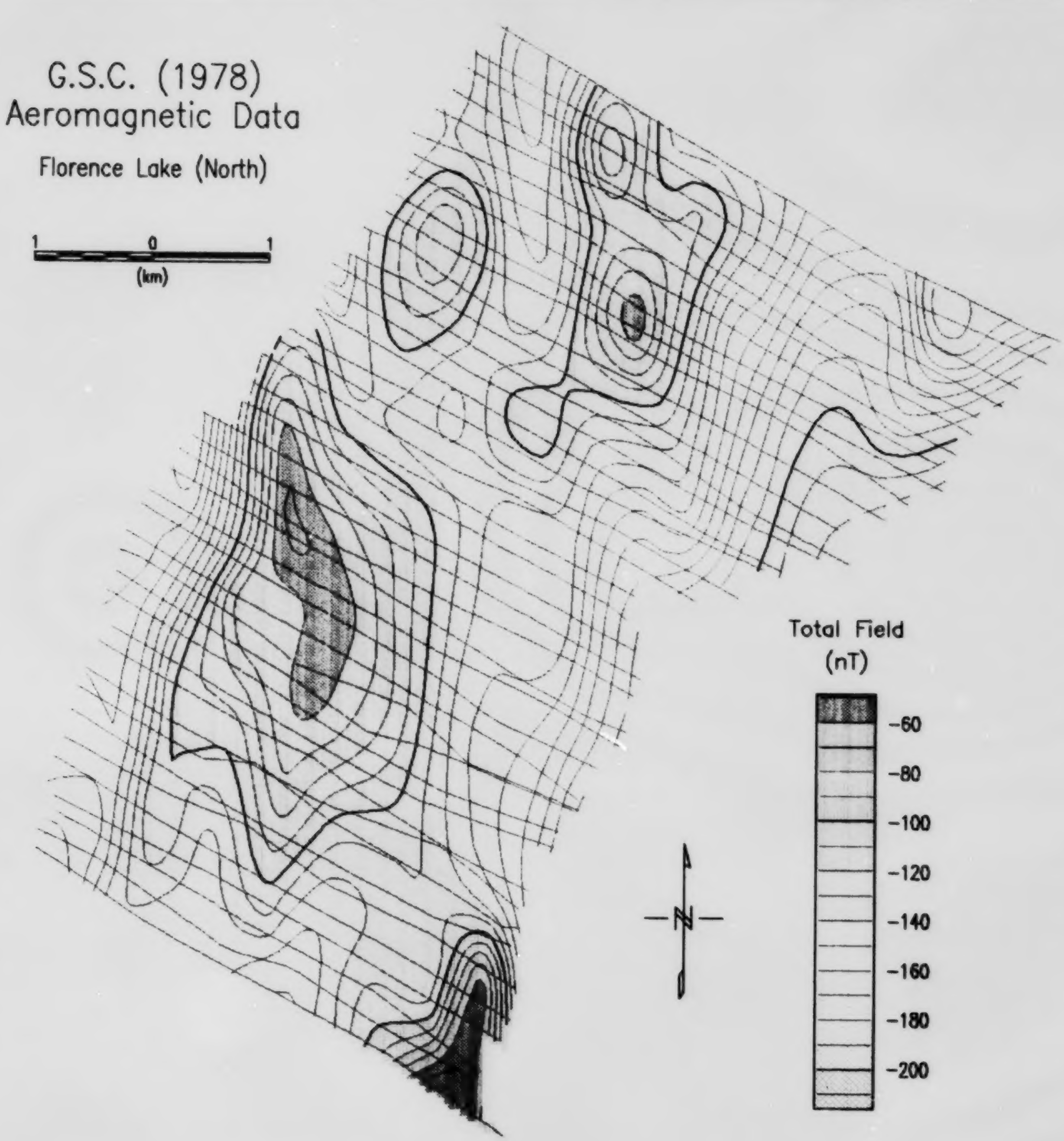
G.S.C. (1978)
Aeromagnetic Data
Florence Lake (North)
1
0
1
(km)
Total Field
(nT)
-60
-80
-100
-120
-140
-160
-180
-200
N

Figure 4. *Shaded, aeromagnetic contour map for the same area as in Figure 3, produced from the existing regional dataset flown by the Geological Survey of Canada during 1978. Note in the legend that total-field magnetic data have been referenced to the IGRF and are variably shaded, similarly to Figure 3, to indicate areas of anomalous magnetic field intensity. The survey was flown at an altitude of 300 m and average flight-line spacing of 800 m.*

linear features are much better defined on Figure 3, and, in certain areas, broad anomalies on Figure 4 are clearly defined as two distinct anomalies on Figure 3. A similar comparison of colour, shaded-relief images from the two datasets shows this contrast in detail more clearly. The superior resolution of features in the industry survey can be largely attributed to the lower survey altitude and more closely spaced flight lines. The east-southeast orientation of flight lines in the industry survey (in contrast to east-west oriented flight lines in the regional survey), perpendicular to the strike of ultramafic units in the Florence Lake area, is also a minor contributing factor to the difference in resolution of features in the two surveys.

Through digital processing and presentation, the example from the Florence Lake area has clearly shown the advantages of digitally capturing geophysical data from industry surveys, presently existing as maps in mineral-assessment reports. The geophysical data component of existing assessment reports for the province represents a huge volume of information. If the bulk of this information could be obtained from the sponsoring companies as raw digital data (as magnetic tapes or other storage media), the savings in resource requirements would be substantial. As well, much of the original information recorded during the survey is simply not recoverable from maps submitted with assessment reports. For example, subtle information, which may have been recorded at locations between contours on a contour map, cannot be retrieved. As well, certain anomaly maps and interpretation maps only provide the locations of the most anomalous data values or those that the contractor or interpreter considered to be significant. Requests for copies of any digital data have been sent to the mineral-exploration companies, which sponsored all airborne geophysical surveys in Labrador.

Digital airborne geophysical data from 52 survey blocks were submitted on magnetic tapes by Labrador Mining and Exploration Company Limited (1980, 1982). To evaluate the advantages of having original digital survey data, profiles of total field magnetic data for an area approximately 75 km southeast of Schefferville (Block 135 South) were extracted from this dataset. The data were then gridded to a 40-m grid-cell size, approximately one fifth the nominal flight line spacing of 200 m. Images generated from the gridded datafile showed the existence of inter-flight-line levelling errors and data 'spikes' or erroneous values distributed randomly throughout the data. The data spikes were removed by applying a non-linear filter to the profile data and levelling errors were diminished by adding constants to the magnetic values along certain profiles and by directional filtering applied to the gridded datafile. A shaded, contour plot of the corrected data is shown in Figure 5. The series of accentuated magnetic linear features that strike to the north in the centre of this map and swing off to a north-northwest direction near the top of the map correspond to the locations of iron formation.

As in the Florence Lake example, a corresponding window of gridded data values was extracted from the (200-m grid cell) GSC gridded data for comparison and evaluation. Figure 6 is a shaded contour plot from the GSC data for the area covered by Figure 5. Although all of the larger anomalies that exist on Figure 5 are present on Figure 6, these anomalies are broader and less defined. In the western half of the map area, the region underlain by iron formation, two distinct linear anomalies, evident on Figure 5, are represented as a single broad anomaly in Figure 6. Thus, although the regional trends and larger magnetic features are apparent in the GSC data, the detailed (digital) survey data contains sufficient information to distinguish anomalous zones to much finer resolution.

When compared to the Florence Lake area, the availability of raw digital data for Block 135 (South) significantly reduced the time required to prepare, process and output a corrected, gridded dataset. The entire process, from retrieving the raw ASCII data on magnetic tape to the levelled and filtered grid file, required approximately one day for an operator using a PC. In addition, having established a processing sequence for the first of the 52 areas for which digital data is available, the process could be automated to reduce operator intervention and the total processing time reduced to half. Whereas the optimum method (digitizing data plots, at present) of data input requires time on the order of weeks to months per map area, the time required to process data already held in digital for an equivalent area is on the order of hours to days. Thus, not only are there distinct advantages, in terms of data detail and accuracy, of obtaining the original data in digital from source, but significant savings in resource requirements can be realized.

RESULTS AND CONSIDERATIONS FOR THE FUTURE

The digital index of airborne geophysical surveys for Labrador will be expanded to include coverage of the larger geophysical (and possibly geochemical) surveys conducted on the ground. Upon completion of the Labrador digital index, a similar index will be constructed for insular Newfoundland. Since coordinates for survey outlines in the Labrador index will be in UTM zone 20 and those for insular Newfoundland will be referenced to UTM zone 21, the present plans are to maintain a separate index for the two parts of the province.

As computer technology evolves (increasing data storage, internal memory and scanner resolving capacities; advances in processing software), the capability to effectively edit large vector datasets, such as the scanned aeromagnetic contour map for Florence Lake, may soon become realistic on a PC. If the software can be adapted to automatically extract flight-line/contour intersections from such vector files, then digital scanning may prove to be a more cost-effective alternative to manually digitizing the data. An advantage of the digital scanning method is that a raster image file is generated as an intermediate product. This file can be archived for future editing and viewing within image analysis systems. At present, however, such advantages as the selective entry of data and drastically reduced database sizes afforded by digitizing far outweigh the less restrictive disadvantages, thereby making digitizing the better data input alternative.

A comparison of aeromagnetic data from a variety of sources has revealed that the more detailed data from surveys conducted by industry for mineral assessment have the potential for better definition of geological features than the

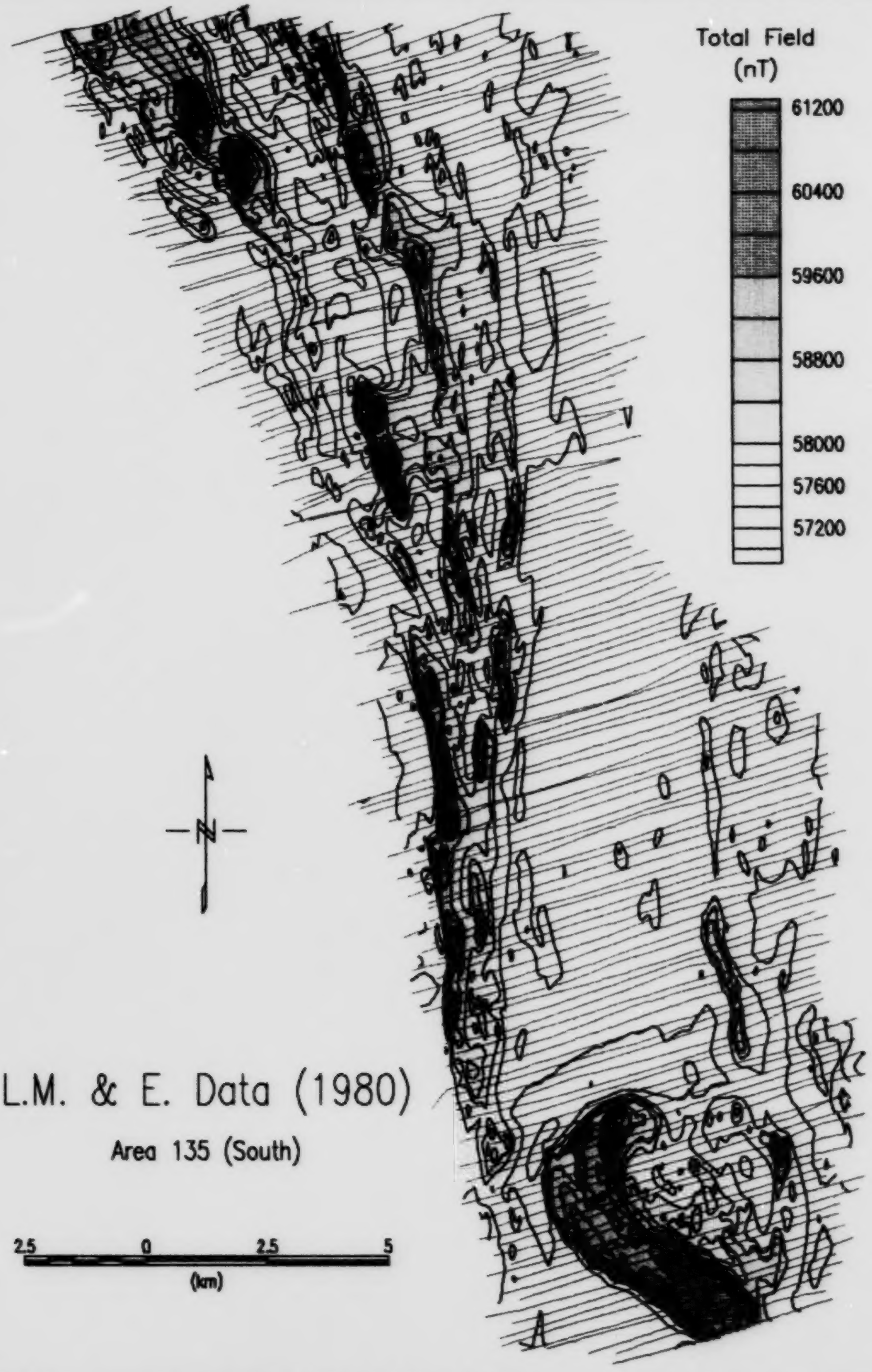
Total Field
(nT)
61200
60400
59600
58800
58000
57600
57200
N
L.M. & E. Data (1980)
Area 135 (South)
2.5
0
2.5
5
(km)

Figure 5. *Shaded, aeromagnetic contour map for Block 135 (South), produced from digital data from surveys flown by Labrador Mining and Exploration Limited during 1980. Total-field values have not been referenced to the IGRF, and are variably shaded to indicate anomalous magnetic field intensity. The diagonal lines striking east-northeast through the area are the survey flight paths. The survey was flown at an altitude of 46 m and average flight-line spacing of 200 m.*

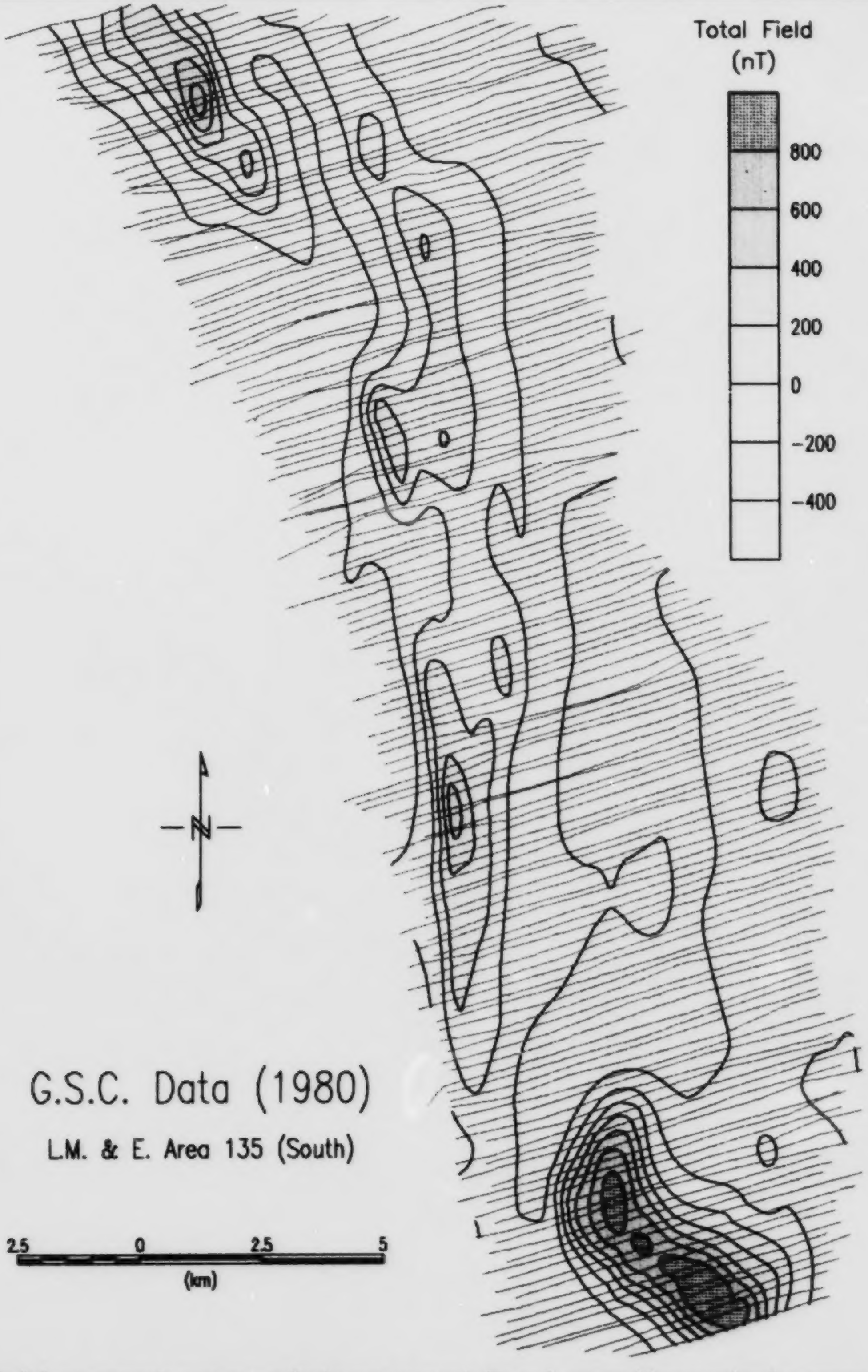
Total Field
(nT)
800
600
400
200
0
-200
-400
G.S.C. Data (1980)
L.M. & E. Area 135 (South)
2.5
0
2.5
5
(km)
N

Figure 6. *Shaded, aeromagnetic contour map for the same area as in Figure 5, extracted from a regional aeromagnetic dataset flown during 1980 by the Geological Survey of Canada. Note on the legend that total-field magnetic data have been referenced to the IGRF. Variable shading to accentuate anomalous magnetic field intensity is comparable to that for Figure 5 and the flight paths from the Labrador Mining and Exploration Limited survey have been included for purposes of cross-referencing. The survey was flown at an altitude of 300 m and average flight-line spacing of 800 m.*

more regional data from surveys conducted by the Geological Survey of Canada. Nonetheless, the complete coverage of regional airborne data provides an excellent framework by which to assess the quality of the more detailed surveys.

Efforts underway to obtain, from source, original digital geophysical data will be further pursued. However, much of the existing data will likely not be available in digital form. Data from older surveys, in particular, would have only been recorded by analog methods and would have to be digitized (or optically scanned) from the analog survey records (if they exist). Due to the volume of existing information in assessment reports, the digital capture of relevant data will be prioritized. Priority will have to be assigned by weighting such factors as existing presentation quality, mineral potential of the area, and active exploration interest.

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SOME USEFUL MAPPING UTILITIES FOR GEOSCIENTIFIC DATA USING A PERSONAL COMPUTER

G. Kilfoil and L. Nolan
Geochemistry and Geophysics Section

ABSTRACT

In recent years, computer programs and algorithms have been developed by the Geological Survey Branch to assist in the processing and presentation of existing geoscientific databases. In an effort to promote the use of the databases held by the Branch, several of the more useful routines are described here. These PC-based utilities range in complexity from a simple coordinate transformation routine to more elaborate software for data processing and display of gridded geochemical and geophysical data.

INTRODUCTION

Several PC-based computer utilities have been developed at the Geological Survey Branch to streamline map production. In most cases, these utilities were originally written to perform a specific task for which no program existed at the Branch. Later, as their day-to-day usage increased, certain utilities were broadened and updated to handle a variety of computing situations or tasks and thereby became more general in application.

The utilities can be divided into two categories: general-purpose mapping utilities and those that perform a particular task with gridded datafiles. A brief description of the function of each utility is listed in Table 1 in the order that they appear in the text.

All utilities are compiled for use on an IBM-compatible personal computer running under the MS-DOS or PC-DOS operating systems. Although the presence of a math co-processor is not strictly required by these programs, the more computation-intensive utilities, such as **DISPLAY** or the gridded-data filtering routines, will operate significantly more slowly in the absence of a math co-processor, particularly if executed on a PC, based on the Intel 8086/8088 or 80286 micro-processors. **GRIDVIEW** and **DISPLAY** require at least a VGA graphics board and colour monitor; all other utilities will work with monochrome or CGA graphics capabilities. Disk storage space is required for input (and output) datafiles for most of the utilities. Although all utilities will function from floppy disk drives, the presence of a hard disk (with a disk cache) in the PC is recommended when executing the gridded-data manipulation routines, as disk access is quite intensive for these utilities.

The source code for all the general purpose utilities was written in Basic and compiled with the Microsoft QuickBASIC optimizing compiler (Microsoft, 1988). In contrast, the utilities that operate on gridded datafiles have been written and compiled in FORTRAN (Microsoft, 1989), due to its superior data-handling capabilities. The general-purpose utilities tend to be smaller and each performs a specific task, but such tasks are generally required more often than those that the gridded data-handling utilities fulfil.

The gridded data-handling utilities can be further classified into general-purpose grid utilities and grid-filtering utilities. Among the tasks performed by the general-purpose grid utilities are: creating VGA colour images from gridded datafiles (in real time), merging gridded datafiles from adjacent areas, and converting gridded datafiles to alternative formats. In contrast, the grid-filtering utilities alter the contents of a gridded datafile by passing a small matrix of filter values over the gridded data elements in order to extract or remove a particular trend or component from the data. The dimensions and formats of filtered grid files remain unchanged relative to the unfiltered versions.

GENERAL-PURPOSE MAPPING UTILITIES:

UTMCON: Converts geographic coordinates of Latitude and Longitude to Eastings and Northings within a user-specified UTM projection, or the reverse.

This handy utility is designed to convert sets of geographic coordinates between the spherical (i.e., Latitude and Longitude) and Universal Transverse Mercator (UTM)

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Table 1. List of documented utilities for mapping geoscientific data and their functions

General Purpose Utilities:	
UTMCON:	Converts coordinates of Latitude and Longitude to Eastings and Northings within a user-specified UTM projection, or the reverse.
UTMCON2:	Converts Easting and Northing coordinates within one UTM zone to equivalents within a second zone.
MAP_CORN:	Automatically determines map area titles and UTM corner coordinates from their NTS designations
UNISTAT:	Provides basic univariate statistics on numerical data sets and generates publication-quality cumulative frequency plots.
GRIDVIEW:	Examines the spatial distribution of data points to aid in selection of gridding parameters.
Grid Utilities (General Purpose):	
DISPLAY:	Interactively displays the contents of binary gridded data files as images on a PC equipped with a VGA graphics card and monitor.
MERGEO:	Merges two binary files of gridded data and/or extracts a rectangular window of the input data for output.
GRD_ASC:	A conversion utility to dump all (or a portion) of a binary grid file to an ASCII file, or vice-versa.
MASK:	Compares two binary gridded data files, and replaces data values within the second with null values at all grid cell locations where null values exist within the first.
GRD_LEG:	Replaces data within a rectantular area of a binary gridded data file with a range of values corresponding to a grid legend.
Grid Utilities (Filtering):	
MEAN:	Applies a Mean filter of variable-width to a binary gridded data file.
MEDIAN:	Applies a Median filter of variable-width to a binary gridded data file.
ATM:	Applies an Adaptive Trim Mean filter of variable-width to a binary gridded data file.
FENCE:	Applies a Fence filter of variable-width to a binary gridded data file.

coordinate systems. The coordinates can be input and/or output either to the computer screen or to a user-specified datafile.

The user is first given the choice of converting Latitudes and Longitudes to UTM Easting and Northing coordinates, or vice-versa. Then the default central meridian, UTM zone and scale factor are displayed and the user is provided the opportunity to change any of these. By choosing a non-standard central meridian and/or scale factor the user can select a specific Transverse Mercator (as opposed to UTM) coordinate system for transformation. If the UTM zone is specified, the central meridian for that zone is taken as the default value.

The user is then queried as to whether input from or output to a file is desired. If either is selected, the name of the file is requested. If input from a file is selected, the first fields of each input line must be UTM Easting, Northing and zone when the UTM conversion option is specified or Latitude and Longitude for spherical coordinate conversion. On output, these fields will be replaced with their equivalents in the new coordinate system and all remaining character data on the input line will be output unchanged.

UTMCON will continue to convert pairs of geographic coordinates until either the end-of-file is reached on input from a data file or zeros are typed as coordinates during keyboard entry.

UTMCON2: Converts Northing and Easting coordinates within one UTM zone to equivalents within a second zone.

UTMCON2 is actually a special version of **UTMCON** that converts coordinates from one UTM zone to another in a single step. This utility has proven useful for converting the coordinates measured from maps in one UTM zone to those for an adjacent zone before the datasets are merged to produce a regional map.

The user is queried for the UTM zones to be used on input and output. As in **UTMCON,** the input data source can be selected either as a datafile or the keyboard; output data destination can be specified as a datafile or the screen.

```
--- Program to Calculate UTM's for Corners of NTS Map Sheets ---

Do you want printer output? n
Do you want to output polygons of corner coordinates to a file? n
Do you want to display corner coordinate windows on the screen? n

Do you want to round map corner extremes (UTM) to a nearest factor? y
Enter a rounding factor for the extremes of UTM corners (e.g. 100)? 100

(1) Central Meridian: 57 ; (2) UTM Zone: 21 ; (3) Scale Factor: .9996
To Change a Parameter, Type Number in Brackets; Otherwise Hit <Enter>?

Input NTS Map: e.g., 'NTS 12A'; '12A' (0 to quit; -1 to alter params)? nts 2d/4

      Map Title:   Twillick Brook
     Map Number:   2D/4
          Scale:   1:50 000
 Latitude range:   48.00 to 48.25 degrees North
Longitude range:   56.00 to 55.50 degrees West

     (48.25, 56.00)                                      (48.25, 55.50)
  ( 574236.32,5344351.72)    -----------------------    ( 611353.68,5344955.94)

            |                                                  |
            |                                                  |
            |                                                  |

     (48.00, 56.00)                                      (48.00, 55.50)
  ( 574597.46,5316565.14)    -----------------------    ( 611895.46,5317169.95)

Rounded UTM Window of Map Corner Extremes:
  ( 574200.00,5316500.00)    -----------------------    ( 611900.00,5345000.00)

Inputs NTS Map: e.g. 'NTS 12A'; '12A' (0 to quit; -1 to alter params)? 0
```

Figure 1. *Example use of* **MAP__CORN** *showing results for the Twillick Brook map area, NTS 2D/4, central Newfoundland. (User responses are in bolder text).*

MAP__CORN: Automatically determines map-area titles and UTM corner coordinates from their NTS designations.

This utility provides the title and corner coordinates for a map area by simply referring to the 1:250 000- or 1:50 000-scale map sheet by its (Canadian) National Topographic System (NTS) reference. **MAP__CORN** is a hybrid version of **UTMCON** that determines and reports the four corner coordinates of the map area in both spherical and Easting/Northing of a user-specified UTM zone.

MAP__CORN searches a list of map-area titles in a DOS file 'MAPS.LST', which should be located in the current working directory. If a file by this name does not exist in the current directory, the user will be prompted for the path to its location. Presently, this list contains only the map-area titles for the area covered by the Province of Newfoundland and Labrador, but the list may be expanded or replaced with a list for any other province/territory or for the whole of Canada. In practice, the map list should be kept to a minimum as the response time for **MAP__CORN** may be slowed by searches through for very long lists of map titles. If a map title for a particular NTS designation is not located within 'MAPS.LST', the user will be prompted for the map title.

Individual 1:50 000-scale map areas are distinguished from 1:250 000-scale map areas by including the slash character (/) in the NTS designation. For example, the 1:50 000-scale map area for Twillick Brook in central Newfoundland would be specified by typing 'NTS 2D/4' (or simply '2D/4'), whereas the 1:250 000-scale map for Gander Lake (which contains the Twillick Brook map sheet) would be specified by typing 'NTS 2D'. Figure 1 shows an example session in which the corner coordinates for the Twillick Brook map area have been calculated within UTM zone 21. Note that **MAP__CORN** iterates the calculation of corner coordinates for additional map areas until a '0' is typed at the prompt for the next map area.

Additional features of **MAP__CORN** are: the ability to specify the central meridian and scale factor of a specific Transverse Mercator projection; an option to calculate the smallest UTM coordinate window, rounded to the nearest multiple of a factor that contains the all four corners of the map area; an option to output (to a user-specified ASCII datafile) the map title followed by five Easting/Northing data pairs defining a closed polygon of the map-area corner coordinates; and optional output to the default printer. The rounded UTM coordinate window option is useful for determining the range of Easting and Northing coordinates for a particular map area when gridding or windowing (see description of **MERGEO** below) a large dataset to a particular cell size. For example, before interpolating a dataset for the Twillick Brook map area to a 100-m grid cell, a user may wish to execute **MAP__CORN** as in Figure 1 (specifying the 100-m rounding factor) in order to determine the Easting and Northing coordinate ranges that would limit the output grid size to this map area.

UNISTAT: Provides basic univariate statistics on numerical datasets and publication-quality cumulative frequency plots.

UNISTAT (UNIvariate STATistics) is an interactive computer program for the IBM-PC and closely compatible microcomputers designed to assist in the process of analyzing and interpreting geochemical data, element by element (univariate) by providing basic dataset statistics, combined with publication-quality graphical output (Nolan, 1990). Descriptive statistics (minimum, maximum, mean, median, standard deviation and coefficient of variation) are computed using selected options of upper/lower data cut-off limits and logarithmic transformation to produce histograms and cumulative frequency plots. Symmetry, population distributions, range and outliers are readily interpreted from the histogram, and the cumulative frequency plot enables breaks between populations of element concentrations to be uniquely identified. The datafile used by UniStat has a fixed ASCII format (Table 2), which can be created with any text editor or with the file-preparation module, which will allow the user to interactively label variables and flag missing cases in a data matrix. Datafiles created by this module or a text editor can contain up to 40 variables (columns) and 10,000 samples (rows).

Table 2. Example data set showing **UNISTAT** ASCII file format.

```
Title Line–Test Data set
3
Arsenic          ppm
Copper           ppm          –3 variables (cols)
Nickel           ppm
  179.0      6      11
  811.0      3       4
   37.2      5       6        –473 samples (rows)
   34.1      7       7
     .       .       .
     .       .       .
     .       .       .
```

The histogram provides information at a glance on the frequency distribution of the data. It determines if the frequency distribution is symmetrical (normal) or skewed, unimodal (one population) or polymodal (several populations) and whether outliers are present, and also the range of the data.

The cumulative frequency plot provides a graphical presentation of the frequency distribution (cfd) on a normal probability scale. A straight line plot on a normal probability scale indicates a dataset with a single, normal population. If the line is curved or plots as several straight lines, a non-normal distribution or multiple populations in the data is indicated. The points of inflection on the plot can be read off the concentration axis and used as distinct population breaks for plotting geochemical element distributions on contour maps (Sinclair, 1983).

GRIDVIEW: Examines the spatial distribution of data points to aid in the selection of gridding parameters.

The presentation of point observations as a continuous surface for display as a raster image, or for contouring, requires that the irregularly distributed values be transformed into a regular grid. The extent to which this computed grid represents the original data points is controlled by several parameters that must be selected by analysis of the dataset. The grid cell size is the most important parameter that controls the quality of the output grid. A cell size too large will result in the grid being overly smooth with the loss of fine detail and one too small will result in increased processing (not critical) with no improvement in grid detail.

GRIDVIEW displays the set of data points on an x-y grid of user-specified grid cell size, on which a circular search area with a user-specified radius can be displayed. This search area can be moved over the grid allowing visual analysis of sample density with respect to the grid cells to aid in the selection of gridding parameters. The cell size and search area radius can be interactively changed. As well, the circular search area may be moved around the dataset while calculating the points that fall within. A histogram of the number of data points per cell is displayed for each selected grid cell size. An efficient grid cell size is one where the number of cells containing one point is a maximum while keeping the overall grid within a reasonable size for processing.

GRIDVIEW has been designed as an interactive tool to aid in the selection cell size and search radius for gridding-point sample datasets by combining visual displays of the data with informative histograms. Other point data-analysis techniques such as variograms give an indication of the spatial continuity of data-point measurements over distance, which helps determine the search radius parameter but not grid cell sizes or search methods (nearest neighbour, octant, etc.). Variograms are more complex to apply, and typically depend on the data behaving in certain consistent, mathematically describable ways, which is often not the case.

Grid Utilities (General Purpose):

DISPLAY: Interactively displays the contents of binary-gridded datafiles as images on a PC equipped with a VGA graphics card and monitor.

The operation of program DISPLAY, complete with examples of colour and grey-scale images, has been outlined previously (Kilfoil, 1991a). The more comprehensive description contained in the User's Manual (Kilfoil, 1990) has been supplemented by upgrade notes as the utility was debugged and retrofitted with more functions. Only those recent major revisions that have added options to **DISPLAY** are outlined here.

Aside from minor bug fixes, two functions have been added: the ability to produce monotone images in shades of one of several colours, and the capability of overlaying vector files on the images for geo-referencing purposes.

In previous versions of **DISPLAY**, monotone shaded-relief or total-field images could only be generated in 64 shades of grey. Monotone images can now be generated in any one of the additional colours: red, green, blue, cyan, magenta, yellow or grey. The monotone colour appears as an extra option in the **DISPLAY** menu. The monotone colour can be selected by typing 'G' at the menu prompt.

The major drawback to previous versions of **DISPLAY** was the inability to overlay information that would provide geo-referencing on images generated. Presently, **DISPLAY** allows only vector files in Tydac SPANS format to be plotted over images for reference. Vector files consist merely of a series of points defined by their X-Y locations. These points are grouped in the vector file into line segments that define a geographic feature, such as a lake, geological boundary, survey line or a boundary of a map sheet or claim block. Appendix A describes the structure of a Tydac SPANS vector file in more detail.

An extra option line was added to **DISPLAY**'s menu screen to allow the user to toggle the vector overlays off or on by typing 'V'. The colour used in plotting the overlay vectors can be selected from the prompt when vector overlays are turned on. Any of the monotone colours listed above, as well as black and white, can be selected independently for each vector file. Thus, the user may overlay one vector file, such as a file of coastline, in grey and another vector file, such as map boundaries, in black on a colour image. The vector overlays are not stored in memory when the user switches between the image display screen and the menu screen. Instead, the vector overlays are plotted after the image(s) have been generated, each time that the image screen is selected. The user may wish to temporarily turn the vector display off when selecting optimum images for two reasons: the vector overlays plot over the image(s), masking the image information contained beneath and thereby detracting from the image(s) and the plotting of vector overlays can be time consuming, particularly if the vector files are quite large. For both of these reasons, the user is advised to keep vector overlays to a minimum; use only sufficient information to properly geo-reference images generated.

Previous versions of **DISPLAY** contained errors that impaired the operation of the program. In particular, an annoying programming bug, which affected the refresh of the right-hand image in the two-image display mode, often resulted in a partially blank image screen. Several bugs have been corrected in Version 2.0.

Since the ability to overlay geo-referencing information represents a major revision to program **DISPLAY**, the program has been released as Version 2.0.

MERGEO: Merges two binary files of gridded data and/or extracts a rectangular window of the input data for output.

The utility, **MERGEO**, was originally designed to merge the contents of two input binary-gridded datafiles from contiguous map areas, resulting in a larger binary-gridded datafile as output. The utility was later modified to allow a window (defined by a range of geographic coordinates) of the merged dataset to be selected for output.

The binary grid-file type can be selected separately for the input and output datafiles. Currently, Geosoft (Versions I and II) and Geopak binary grid-file formats are supported. Thus, by specifying a different grid-file type for input than for output, **MERGEO** can be used as a data conversion utility.

In addition to the output file, which is specified on the command line, **MERGEO** allows the user the option of specifying either one or two input file names. If only one input file name exists on the command line, no data merging is performed and either the entire contents or that within a data window of the input file will be transferred to the output file. If two input files are specified, **MERGEO** merges data values from the two and outputs the result (or a portion within the specified window) to the output file. **MERGEO** uses the geographic coordinates of the grid origin, stored in the grid header line, to determine the relative positions of data from the two input files and the location of the optional window selected for output. Data values in areas of overlap on the two input grids are averaged before output.

The output data window is specified on the command line by the two: -x=#,# and -y=#,# flags, where x and y indicate respective grid dimensions and #,# specifies the minimum and maximum extremes of the data window expressed in geographic coordinates. Alternatively, the output data window may be specified in grid cells by using the -nx=#,# and -ny=#,# flags. If a coordinate of the specified data window falls outside the coordinate extremes of the input or merged gridded data, that window coordinate is set to the respective data extreme.

Several other features have been added to **MERGEO**. Basic statistics (i.e., minimum, maximum) and a frequency histogram are accumulated during the merging process and these are stored in the headers of the output grid files. **MERGEO** will merge and/or convert the formats of data stored in either of the row or column orientations used by Geosoft. A gap (represented by null values) in the input and/or merged datafile may be filled by applying a minimum curvature (Akima, 1972) interpolation from the values of points to either side of the gap. Interpolation (applied along the grid storage direction) can be selected by the command line flag **-s=n**, where **n** specifies the maximum gap width in grid cells.

The user's manual provides further details on the use of the **MERGEO** utility.

GRD_ASC: A conversion utility to dump all (or a portion) of a binary grid file to an ASCII file, or vice-versa.

GRD_ASC is a reformatting utility that provides a two-way link between binary-gridded datafiles and their ASCII equivalent. The routine was originally written in response to requests from those who purchased gridded aeromagnetic data released as digital Open Files and were viewing those files with **DISPLAY**. Several people expressed an interest in having a means to dump the data contents of a Geosoft or Geopak binary-gridded file to ASCII characters, so that data values could be viewed with any ASCII editor or sent to a printer.

During the design of **GRD_ASC**, its functionality was extended to include the ability to import existing datafiles stored in ASCII to a Geosoft or Geopak binary-storage format, so that the data may be viewed with the **DISPLAY** image generating utility. In its present form, **GRD_ASC** may be used to:

- Dump the entire data contents of a Geosoft or Geopak format binary-gridded datafile to ASCII.
- Select a window of a Geosoft or Geopak grid for export to ASCII. This reduces storage requirements and allows the user to quickly view data values only from the particular area of interest.
- Import a datafile, stored in ASCII, of gridded data values to a binary file in Geosoft or Geopak format. The ASCII input file could be data dumped from any other gridding or viewing software.
- Select a portion of an ASCII datafile for import to a binary-gridded datafile, in order to reduce size or processing time. Program **MERGEO** may later be used to piece smaller binary grid files into a large data file.
- Alter most of the descriptive grid parameters that are stored in the header lines of a Geosoft or Geopak binary datafile.

When operated in the data import mode, **GRD_ASC** also performs simple statistics on the selected data window and stores these along with a 32-interval frequency histogram of data values in the header file of the output binary grid file.

The functions and operation of **GRD_ASC** are controlled by a series of command line flags, a simple GXF header command file (in ASCII), or through a combination of both. The command line usage for **GRD_ASC** can be obtained by typing **'GRD_ASC'** at the DOS prompt. By default **GRD_ASC** operates in the data export mode. Users are cautioned that binary-gridded datafiles require considerably less storage space than their ASCII equivalent. Therefore, if data-storage space is limited, only smaller data windows from large binary grid files should be selected for export to ASCII.

The data import mode can be selected on the command line with the **-i** flag. When **GRD_ASC** is used in the import mode, parameters used for control are assigned default values. This enables the utility to be used to create a binary grid file from a generic ASCII datafile with minimal preparation by the user.

The format of all ASCII files exported from or imported to **GRD_ASC** should conform to the Grid eXchange File (GXF) format as described in Appendix G (Revision 1.0) of the Geosoft 2-D Mapping System user's manual (Geosoft, 1990). The GXF format was chosen because of its adaptability to a wide range of existing ASCII data formats. The GXF format was further revised for **GRD_ASC** to include such features as the ability to skip any number of descriptive header text lines in the ASCII datafile, or to skip any number of text lines at the start of each row of data.

The **GRD_ASC** utility is included with all requests for the **DISPLAY** utility. A nine-page, comprehensive user's reference guide (Kilfoil, 1991b) describes the various functions of **GRD_ASC** and the formats of the GXF and binary-gridded datafiles.

MASK: Compares two binary-gridded datafiles and replaces data values within the second with null values at all grid-cell locations where null values exist in the first.

MASK simply masks or blanks out areas in a gridded datafile whose locations correspond to null values in a second gridded datafile. Null values are usually placed in a gridded datafile to designate areas where no data exist. The interpolating routines used to create gridded datafiles often pad the areas outside the input data with null values in order to ensure that a gridded datafile is rectangular.

In the process of grid generation, grid values may be over- or under-estimated when interpolated across areas where no input data exists, such as across inlets, straits, large lakes or recreational areas. If a gridded datafile exists with such areas blanked out, **MASK** can reference null values in these areas to blank out portions of other interpolated data grids.

MASK is compatible with Geosoft or Geopak format gridded datafiles.

GRD_LEG: Replaces data within a rectangular area of a binary-gridded datafile with a range of values corresponding to a grid legend.

This utility is designed to insert a rectangular legend into a Geosoft or Geopak gridded datafile. **GRD_LEG** was originally developed as an addendum to the Geosoft mapping system, as a means of labelling colour maps, produced from gridded datafiles, with a bar-type legend showing the range and scale of data.

The bar-type legends are created by overwriting areas in a gridded datafile with false data values; therefore, the process is irreversible in that any data that previously existed in the area of the legend will be destroyed. Bar-type legends can be placed in the gridded datafile with their long axis oriented horizontally or vertically.

Two types of legends can be generated and inserted into a gridded datafile by **GRD_LEG**: one with discrete cells of a user-specified size or a continuous strip legend having data varying linearly from one end to the other. The data values to be placed in a legend with discrete cells are input from a small (ASCII) control file (equivalent to Geosoft's zone file), one value per line and usually arranged in ascending order. Each value in the control file will produce one cell in the legend to a maximum of 50 cells. In the case of a continuous strip legend, the values placed in the legend will be linearly stretched from the values of the first to the last data entries in the control file, since these are assumed to be the approximate minimum and maximum data values, respectively, for the grid. The size of each cell in the legend is determined by the command line flag **-c=x,y**, where **x** and **y** are sizes expressed in grid cells for the two dimensions. Therefore, the long dimension of a legend (in grid cells) is determined by multiplying the length of each legend cell by the number of values in the control file. If the legend is longer than the corresponding grid dimension, the user is notified and no legend will be created.

The locations of legends within the gridded datafile is controlled by the **-o=x,y** flag, specifying the legend origin (**x** and **y** refer to numbers of grid cells from the grid origin) as a command line option. Legends are usually placed in the gridded datafiles within areas of null values. If a location outside the dimensions of the gridded datafile is specified, the output grid file will be automatically expanded to accommodate the legend. This option is useful in cases where a grid file contains little or no areas of null values.

Specification of an output grid-file name when using **GRD_LEG** is optional. If only one grid-file name (the input file) is included on the command line, the legend will be inserted into the input grid file; if an output file is also named, the contents of the input grid will be written to the output grid file together with the legend, and the input file will remain unchanged.

Gridded Data Filtering

The amount of local variation (including noise) contained in a dataset can be varied during the grid interpolation process by the gridding parameters chosen. Grid appearance and accuracy (to the original data) can be controlled by grid cell size, search radius, neighbourhood selection pattern and interpolation algorithm. Gridded datasets that are destined for a specific purpose, such as contouring, can be created using the initial gridding parameters to produce smooth contours. An alternative method of producing smooth contours from gridded data is to choose initial gridding parameters that will produce an output grid that honours input data points and retains as much of the original information as possible. This detailed grid can be filtered to produce output grid files that can be used for contouring, outlier detection, anomaly detection, smoothing of local variation, edge detection as well as other specialized applications. The aim in most filtering situations is to minimize the background noise while enhancing the data signal.

A set of filter programs were written in FORTRAN-77 on a IBM-PC and are designed to read and write the binary grid-storage formats, which are commonly used by geochemical and geophysical software packages. These programs do not require the computer system to have graphics capabilities but, due to the mathematical calculations involved, a math co-processer is recommended for speed considerations.

MEAN: Applies a Mean filter of variable-width to a binary-gridded datafile.

The most common grid filter is the moving average or rolling mean filter, which calculates a new grid based on the mean of a n by n window for each cell in the detailed grid. There are several variations of the mean filter (low pass, hanning, etc.) but they are all very similar in operation and results. The amount of smoothing is controlled by the size of the window (3 by 3 is common) and the number of times it is passed over the grid. Mean filters diminish localized high/low values to enhance regional patterns in the grid. Mean filters are used to improve the appearance of the display data in cases where the loss of localized information is not critical. Mean filters may be strongly affected by extreme outliers in the data, which in many cases are due to errors.

MEDIAN: Applies a Median filter of variable-width to a binary-gridded datafile.

The median (the midpoint value of a dataset) is replacing the mean (the average of a dataset) as a descriptive statistic that better describes the central tendency of a given set of data values in many cases. Similarly, in grid filtering methods the median and mean filters are commonly used to filter grids to remove spurious values. The median filter does have some advantages over the mean. Edges and sudden changes in data values between grid cells are retained by the median filter, but small scale variation is largely removed.

ATM: Applies an Adaptive Trimmed Mean (ATM) filter of variable-width to a binary-gridded datafile.

The mean and median filters each have common and unique features that are optimum in specific situations. The ATM filter (Chork and Mazzucchelli, 1989) combines features from both techniques to remove background noise, retain edges and sudden changes in data values and strip outliers from the data. As the name suggests, the filter calculates the mean using the median as a robust measure to remove outliers from the active window. Only cell values that fall within approximately 2 standard deviations (computed from the median absolute deviation about the median) of the median are used to calculate the mean for the cell window. The degree of generalization produced can be controlled by increasing the size of the active window area (3 by 3, 5 by 5, 7 by 7, etc.) beginning with the average sample density of the original survey as the smallest window that will have any distinguishable effects. This approach can be quite useful for producing geochemical grids, which emphasize geological structure (Davenport and Nolan, 1991) while still retaining anomalous signatures.

FENCE: Applies a Fence filter of variable-width to a binary-gridded datafile.

The majority of values in a regional geochemical data set comprise the geochemical background due to variations in bedrock lithology, and a minority of these values are anomalies related to mineralization. Depending on the element, this signal-to-noise ratio will vary. For instance, gold data is generally noisy due to the small-scale variance and most samples fall below the analytical detection limit. The filtering routines discussed so far remove the outliers from the dataset, whereas the Fence filter (Chork and Mazzucchelli, 1989) highlights the anomalous regions. Using basic Exploratory Data Analysis (EDA), the sample median of a central block of cells with **n** by **n** area is compared to the upper and lower quartiles of a annulus ring of cells of given radius (fence width) from the central cell being calculated (Figure 2.). In general, the **FENCE** filter measures the strength of a localized anomaly compared to its surroundings. If the median of the central cells is greater than the upper quartile (75th percentile) of the fence cells, based on robust statistical measures of the fence cells, then the central cell is flagged as anomalous. The Fence filter will flag anomalous cells in areas of relative low as well as high data values. The degree of generalization produced by the filter can be altered by varying the size of the central block of cells and the number of cells in the fence area. The cells that fall between the central cells and the fence ring are excluded from the computations.

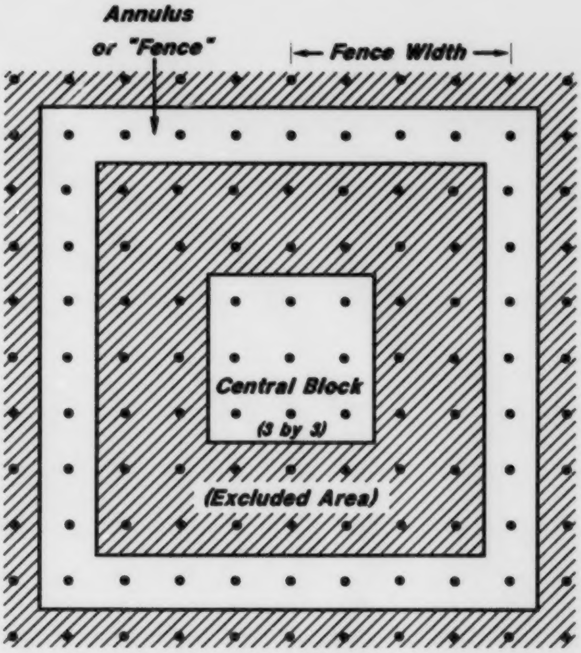

Figure 2. *Schematic diagram of a Fence Filter where dots represent the centres of grid cells. The central block has an effective radius of 1 (3 by 3) and the fence width is 4 grid cells.*

Filtering Procedures

The filtering utilities are implemented and controlled through a command line parameter list of input binary grid file, output binary grid file, file type (Geosoft Version I or II, or Geopak), window size in grid cells and, for the **FENCE** filter, the size of the exclusion fence in grid cells. Each filtering utility will provide a help screen if the filter name is entered without a command line parameter list. Experience with these filters has shown that combinations of different filters may be passed over a grid in succession to obtain the desired results. For example, an **ATM** having a wide window followed by a **MEAN** with a narrow window will result in a output grid that retains large scale features (**ATM**) in the data where the transitional areas between features will be smoothed to improve the appearance (**MEAN**). When choosing the filter window size, consideration should be given to the sample density of the original data. Using a window that is smaller than the original sample density will result in a grid that is very similar to the input grid. The selection of the window size can be aided by using programs such as **GRIDVIEW** to view sample density (smallest window) and variograms to show variation of sample values over distance. However, some experimentation with window widths and filter types may be required before an optimum is reached.

The principles of filtering logic as applied to exploratory data analysis have been discussed by Chork and Mazzucchelli (1989).

CONCLUSION

The functions of several PC-based computer utilities developed at the Geological Survey Branch and adapted for general usage are described above. Many other data-handling utilities have been developed, but their functions are very specific and tailored to the software and hardware that exists within the Branch. These routines have not yet been generalized to handle a variety of computer configurations that may exist in the public domain. As those utilities are updated and as others are developed, their descriptions will be described in future Current Research papers.

The utilities listed in this article can be made available in executable form on a floppy diskette at a nominal cost of reproduction. Requests for copies of many of these utilities will be accompanied by brief documentations describing the function and usage of each.

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Appendix A

The TYDAC SPANS GIS Vector Interchange Format

The TYDAC SPANS Vector Interchange Format requires that the first line of an ASCII vector file consist of the left justified string 'ARCS'. This is followed by each vector line segment, one after another. The first line of each vector line segment representation is a header line, which must have a minimum of four integer parameters: the vector segment number, the number of points (n), the level designation, and the feature code. The header line parameters must be separated from each other by blank spaces. This header line is followed by coordinate pairs (**X** and **Y**), one coordinate pair per line. Thus, each vector line segment requires **n+1** lines in the ASCII file, where **n** is the number of points in the line segment.

The example that follows is a SPANS Vector Interchange Format file containing two closed polygons: the corner coordinates, in metres within UTM Zone 21, for the 1:50 000-scale NTS 2D/4 and 2D/12 map areas, respectively. The comments (within brackets) in this example file are not required.

```
ARCS                          (Indicates a Vector)
      1      5    0  9963     (Segment Header for NTS
                              2D/4 Outline)
574597.46  5316565.14
574236.32  5344351.71         (Last Coordinates in
611353.68  5344955.94         Segment are Equal to the
611895.46  5317169.95         First: Closed Polygon)
574597.46  5316565.14
      2      5    0  9963     (Segment Header for NTS
573873.75  5372139.46         2D/12 Outline)
573509.76  5399928.40
610263.73  5400531.36
610809.77  5372743.08
573873.75  5372139.46
```

The above file could be used by **DISPLAY** to overlay the two map-area outlines on colour images generated from a gridded datafile. To enable **DISPLAY** to overlay vector files properly, the geographic coordinates of points in vector line segments must be consistent with the coordinate system to which the grid file is referenced. For instance, the above vector file would overlay correctly on images generated from the gridded aeromagnetic data for Newfoundland, released as a digital Open File of binary-gridded datafiles (Kilfoil and Bruce, 1990), which were also referenced to the UTM coordinate system in zone 21.

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THE RELEVANCE OF BALTIC SHIELD METALLOGENY TO MINERAL EXPLORATION IN LABRADOR

C.F. Gower
Labrador Mapping Section

ABSTRACT

The hypothesis that Laurentia and Baltica evolved together as a single continental landmass during the mid-Proterozoic is now well established, and has important implications with respect to mineral exploration in Labrador. In Baltica, mining has a 2000-year history, mineral prospecting has been carried out for 500 years, and the geological database has been steadily accumulating over the last 150 years. This vast wealth of accumulated knowledge is thus available for application to mineral exploration in Labrador.

In this article, correlations are restricted to regions south of the Archean cratonic nuclei. The Kalevian Group (ca. 2.10 to 1.94 Ga), which was deposited as a continental marginal prism on the flank of the Karelian cratonic nucleus, and Cu−Zn−Co−(Ni) mineralization within it (e.g., Outokumpu), is compared with rocks deposited on the northwest margin of the Makkovik Province, known to host similar mineralization.

The Svecofennian orogen (1.91 to 1.82 Ga) is equated with the Makkovik−Ketilidian orogen. Inasmuch as only the fringe of the Makkovikian−Ketilidian orogen flanking the Archean cratonic nucleus is preserved in Labrador, only the northernmost section of the Svecofennian orogen is deemed relevant to metallogeny in Labrador. From a metallogenic standpoint, however, this region is the most important part of Baltica. The marginal part of the Archean microcontinent that was reworked during Svecofennian orogenesis is host to Mo−U−(W−Sn) mineralization, south of which are the arc-related Cu−Zn−Au massive sulphide deposits and Cu−Mo−(Au) porphyry-type mineralization of the Skellefte district and then a zone of Ni mineralization. In the Makkovik orogen, Mo−U mineralization is flanked on its eastern side by a concentration of Cu−Pb−(Au) showings, suggesting an analogous situation. In both Baltica and Labrador, at least some of the Mo−U mineralization has been linked with post-orogenic 1.81 to 1.77 Ga granitoid rocks.

The next major accretionary event took place between 1.71 and 1.60 Ga, resulting in the Gothian−Kongsberg orogen in Baltica and the Labrador orogen in Laurentia. Pre-orogenic pelitic and psammitic schists in Baltica (ca. 1.76 Ga) are host to stratiform Cu−(Co) deposits, which have been compared to Besshi-type ores. Similar, but barely prospected, sulphide-impregnated, locally Cu-mineralized zones, are common in parts of southern Labrador. The later, syn-orogenic Gothian−Kongsbergian rocks appear devoid of related mineralization, except in felsic volcanic rocks deposited on older crust at the eastern fringe of the orogen, hence the only Laurentian counterparts judged to hold prospecting potential are analogous felsic volcanic rocks at the northern margin of the Labrador orogen. Coeval (1.64 Ga) rapakivi granitoid rocks in Finland host Sn−Be−W−Pb−Zn deposits, however, and comparable mineralization could exist in the southern part of the poorly known 1.64 Ga Mealy Mountains Intrusive Suite in Labrador, if uppermost intrusive levels are preserved.

Following crustal stabilization between 1.59 and 1.51 Ga, both Baltica and Laurentia experienced anorogenic conditions, lasting until 1.27 Ga. Apart from V−Ti deposits associated with mafic intrusions, mineralizing activity during this period appears to have been meagre in Baltica. Such need not be true of Labrador, where there is sufficient contrast in structural setting of a wide range of younger anorogenic rocks to allow considerable promise for both Sn−Be−W−Pb−Zn granophile mineralization and Ti−V mafic-intrusion-related mineralization.

The final stage in the evolution of mid-Proterozoic Laurentia−Baltica was the 1.27 to 0.95 Ga Sveconorwegian−Grenvillian orogenic cycle, which embraced several separate tectonic events. Related mineral deposits include Cu−Mo−Au mineralization hosted by rift-related supracrustal rocks, Mo mineralization linked with syn- to late-tectonic granites, Au -Cu−(Pb−Zn−Ag) mineralization associated with late- to posttectonic veins, and Ti−Fe−Ni mineralization in anorthosite. Magmatic Ti−Fe−Ni segregation deposits in anorthosite have obvious analogues in eastern Laurentia (e.g., Lac Tio, Quebec), but the potential for Grenvillian analogues for the other types of mineralization remains poorly known. Much of the Cu−Mo−Au, Mo, and Au−Cu mineralization has been linked to late Sveconorwegian granitoid rocks. The existence of comparable granitic rocks has only been recognized in recent years in Labrador, and most are likely to occur in yet-to-be-mapped areas in southernmost Labrador. These granites offer a promising and completely untested target for mineralization, and, following the completion of mapping in the next few years, an up-to-date geological database will be available for focussed exploration.

INTRODUCTION

> *From 1500 onwards there has been systematic prospecting in Sweden, undertaken by the state, private individuals and companies.* (Grip, 1978).
>
> *Relatively little mineral exploration has been carried out in Labrador ... and large tracts remain virtually unexplored.* (Swinden *et al.*, 1991).

Laurentia and Baltica formed a single continental landmass during Proterozoic times, following the amalgamation of Archean continental nuclei between 1.90 and 1.80 Ga (cf. Gower *et al.*, 1990). This hypothesis is now well established and is used in this article as a basis for examining the Proterozoic metallogenic history of the Baltic Shield in an attempt to gain insights regarding how such knowledge could be applied to mineral exploration in Labrador. Conversely, it is also hoped that the Laurentian perspective presented here will be of use to future metallogenic studies in Baltica.

Why study the Baltic Shield at all? Surely, metallogenic comparisons between Labrador and other parts of Laurentia are more pertinent, especially as geological continuity can be demonstrated much more easily. Certainly, rock units do not stop at provincial borders (despite the way their distribution is commonly depicted on many geological maps), and it is obviously important to pay attention to ongoing research and exploration in immediately neighbouring regions. Nevertheless, there are several advantages to be gained by considering metallogenesis in the Baltic Shield that cannot be gained within Canada. Baltica has very long traditions of mining, mineral exploration and geological study. Mining in Baltica has a history extending back over 2,000 years, serious mineral exploration has been carried out for over 400 years, and systematic geological mapping has been in progress for 150 years. This has resulted in the accumulation of a very large metallogenic data base, unequalled in most of Laurentia.

The major theme that underlies this report is that, given the validity of tectonic correlations between Baltica and Labrador, then the huge metallogenic data base for Baltica can be used as a tool for the identification of analogous metallogenic environments in Labrador. The secondary theme is that, despite investigations over such a long period (or perhaps because of them?), new metallogenic environments are still being discovered in the Nordic countries. Apart from demonstrating that success does not come easily, it raises the question, 'what might a region such as Labrador, which is almost totally unexplored by comparison, have to offer?'.

A SUMMARY OF LAURENTIA–BALTICA CORRELATIONS

The Baltic Shield occupies about 1,000,000 km^2 in Sweden, Finland, Norway, the Baltic states and westernmost Russia (Figure 1). It can be broadly divided into three major regions, namely the Archean and associated provinces in the north, the Early Proterozoic Svecofennian orogen in the centre, and the Middle and Late Proterozoic Southwest Scandinavian Province in the southwest. Baltica also extends under Paleozoic platform rocks southeast of the Baltic Sea, and, in the west, forms the basement to the Paleozoic Caledonian orogen.

Although correlations between Laurentia and Baltica exist both north and south of the Archean cratonic nucleus of the North Atlantic–Karelia cratons, attention here is restricted to the regions south of the cratonic nuclei (Figure 1). The major correlations are as listed below. In subsequent sections, key metallogenic features are outlined below for each of the major tectonic groupings in Baltica and potential correlative environments in Laurentia are proposed.

(i) Pre-Svecofennian, Karelian continental margin prism rocks of the 2.10 to 1.94 Ga Kalevian Group are correlated with similar rocks at the northwest margin of the Makkovik–Ketilidian orogen of Laurentia.

(ii) The 1.91 to 1.82 Ga rocks of the Svecofennian orogen are correlated with rocks of similar age within the Makkovik–Ketilidian orogen.

(iii) The 1.81 to 1.77 Ga late- to post-Svecofennian units are correlated with late- to post-Makkovikian–Ketilidian rocks in southern Greenland and Labrador.

(iv) The pre-Gothian–Kongsbergian Stora Le-Marstrand Formation and correlative rocks (ca. 1.76 Ga) are equated with pre-Labradorian sedimentary rocks in the Grenville Province.

(v) The 1.71 to 1.50 Ga Gothian–Kongsbergian orogen is equated with the Labrador orogen that transects southern Labrador.

(vi) Various anorogenic AMCG (anorthosite–mangerite–charnockite–granite) suites having ages between 1.65 and 1.27 Ga in Baltica are compared with similar rocks emplaced within the same time interval in eastern Laurentia.

(vii) Pre-Sveconorwegian and Sveconorwegian rocks having ages between 1.27 and 0.95 Ga are equated with broadly coeval pre-Grenvillian and Grenvillian rocks in eastern Laurentia.

KALEVIAN–MAKKOVIKIAN MARGIN

GEOLOGICAL SETTING

The boundary between the Archean craton and the northern margin of the Svecofennian orogen cannot be drawn precisely, as the Archean rocks to the northeast were remobilized during Svecofennian orogenesis. The Archean margin can be identified, however, on the basis of Sm–Nd isotopic evidence and also by contrasting rock types present north and south of the Raahe–Ladoga zone, with which the isotopic boundary is approximately coincident (Figure 2). The Raahe–Ladoga zone, characterized by wrench faults and an associated negative gravity anomaly, has a movement history spanning 1.90 to 1.75 Ga (Park, 1991). It is, undoubtedly, in economic terms, the most important lineament in the Baltic Shield.

The region north of the Archean–Proterozoic crustal boundary is underlain by Lower Proterozoic supracrustal formations deposited on, and adjacent to, the Archean microcontinent (Karelian continent). Stratigraphic relationships between the various supracrustal suites are complex (Park, 1991), but a key event near the end of this period, from the perspective of this article, was the injection of northwest-trending mafic dykes associated with rifting between 2.2 and 2.0 Ga (Skiöld, 1986: Gaál and Gorbatschev, 1987). Following this event, a passive continental margin developed and a continental marginal prism (Kalevian Group; 2.1 to 1.94 Ga) was deposited. The Kalevian Group grades from proximal, shelf clastic sediments in the northeast to deep-water, distal turbiditic sediments associated with 1.96 Ga slices of ophiolite farther southwest (Gaál and Gorbatschev, 1987; Pharaoh and Brewer, 1990).

METALLOGENY

The most important metallogenic aspect of the Lower Proterozoic rocks immediately north of the Svecofennian orogen is rift-related Cu–Zn–Co–(Ni) mineralization hosted by Kalevian supracrustal rocks. The most famous example is the Outokumpu district, which includes the Keretti, Vuonos, Luikonlahti and Kylylahti deposits (these are not separately indicated on Figure 2). To date, 40 million tonnes of ore have been mined in the district, of which 28 million tonnes have come from the Keretti mine (Parkkinen, 1986). Several additional prospects are known in the Outokumpu district, and in the Kainuu schist belt to the northwest (e.g., Jormua). These are associated with a metal-rich black shale containing 0.14 percent Cu and 0.53 percent Zn (Loukola-Ruskeeniemi *et al.*, 1991).

The mineralization occurs as massive sulphide deposits in a rock association consisting of serpentinite, dolomite, calc-silicate rocks, quartzite, graphitic schist and mica schists. The mineralization is associated with black schists and cherts that were deposited as marine argillaceous sediments between 2.10 Ga and 1.96 Ga. The ore minerals are pyrrhotite, pyrite, chalcopyrite, sphalerite and pentlandite. The average grade of the Keretti deposit is 3.8 percent Cu, 1.0 percent Zn, 0.24 percent Co and 0.12 percent Ni. The ore and its carbon-bearing chemically precipitated host rocks are interpreted to have formed as a result of marine exhalative processes, probably in a rifted continental-margin basin environment (cf. Parkkinen, 1986; Kontinen, 1987; Pharoah and Brewer, 1990).

LAURENTIAN ANALOGUES (2.20 to 1.95 Ga)

The Laurentian analogues of the Kalevian Group are the Moran Lake and Lower Aillik groups in Labrador (Figure 3) and the Vallen and Sortis groups (Allaart, 1976) in southern Greenland. Although none of these rocks is yet precisely dated, it is likely that the rocks formed on the flank of a marginal basin during a ca. 2.20 to 2.00 Ga rifting event. Summaries of the Moran Lake and Lower Aillik groups are given by Wardle and Bailey (1981) and Ryan (1984), and aspects of the metallogeny of the Moran Lake Group have been addressed by Wilton *et al.* (1988) and North and Wilton (1988). The latter two publications identify two types of mineralization in the Moran Lake Group, namely epigenetic Pb–Zn–Cu occurrences at the basal unconformity with Archean rocks, and syngenetic massive Zn–Cu sulphide occurrences within shales of the Warren Creek Formation (cf. also North and Wilton, *in press*). The distribution of mineral occurrences in the Moran Lake Group in Figure 3 is based on Ryan (1984); although Cu mineralization seems to be most ubiquitous, Pb–Zn mineralization dominates at all the major showings.

The shale-hosted mineralization is clearly a close analogue of that found in the Kalevian Group, not only in terms of overall tectonic setting, but also in the nature of the host rocks and the type of mineralization. Some differences between the two areas are that the Outokumpu deposits occur in allochthonous thrust slices, have some associated Ni mineralization, and have been conceptually linked with ophiolites (Kontinen, 1987). Although a simple ophiolite interpretation is not without problems (Gaál and Gorbatschev, 1987), the overall differences suggest a somewhat more distal (from craton) environment than that of the Moran Lake Group. The Lower Aillik Group, farther east, may provide a better analogue, and is known to host sulphide mineralization in graphitic black shales within a dominantly mafic volcanic sequence (cf. Swinden *et al.*, 1991).

SVECOFENNIAN–MAKKOVIKIAN

> *'... it was not until the eighteenth century that exploration activity was reported in the area, now called the Skellefte district. However, it took another 200 years before prospecting was successful.'* (Weihed *et al.*, *in press*).

GEOLOGICAL SETTING

The Svecofennian orogen is divided into three parts (Gaál and Gorbatschev, 1987). These are a northern magmatic arc (Skellefte district), a central intra-arc basin (Bothnian Basin) and a southern magmatic arc (Bergslagen district). Only the northern magmatic arc is deemed here to be of relevance to correlations with eastern Laurentia. The base-metal deposits associated with felsic volcanic rocks in the Bothnian Basin (Hallberg, 1989), and those related to supracrustal rocks in the famous Bergslagen mining district and southern Finland (Lundström and Papunen, 1986; Baker and Hellingwerf, 1988) in the southern Svecofennian will not be reviewed.

The magmatic arc south of the Archean–Proterozoic crustal boundary formed between 1.91 and 1.86 Ga. In the Skellefte district (Figure 4), it consists of a lower submarine,

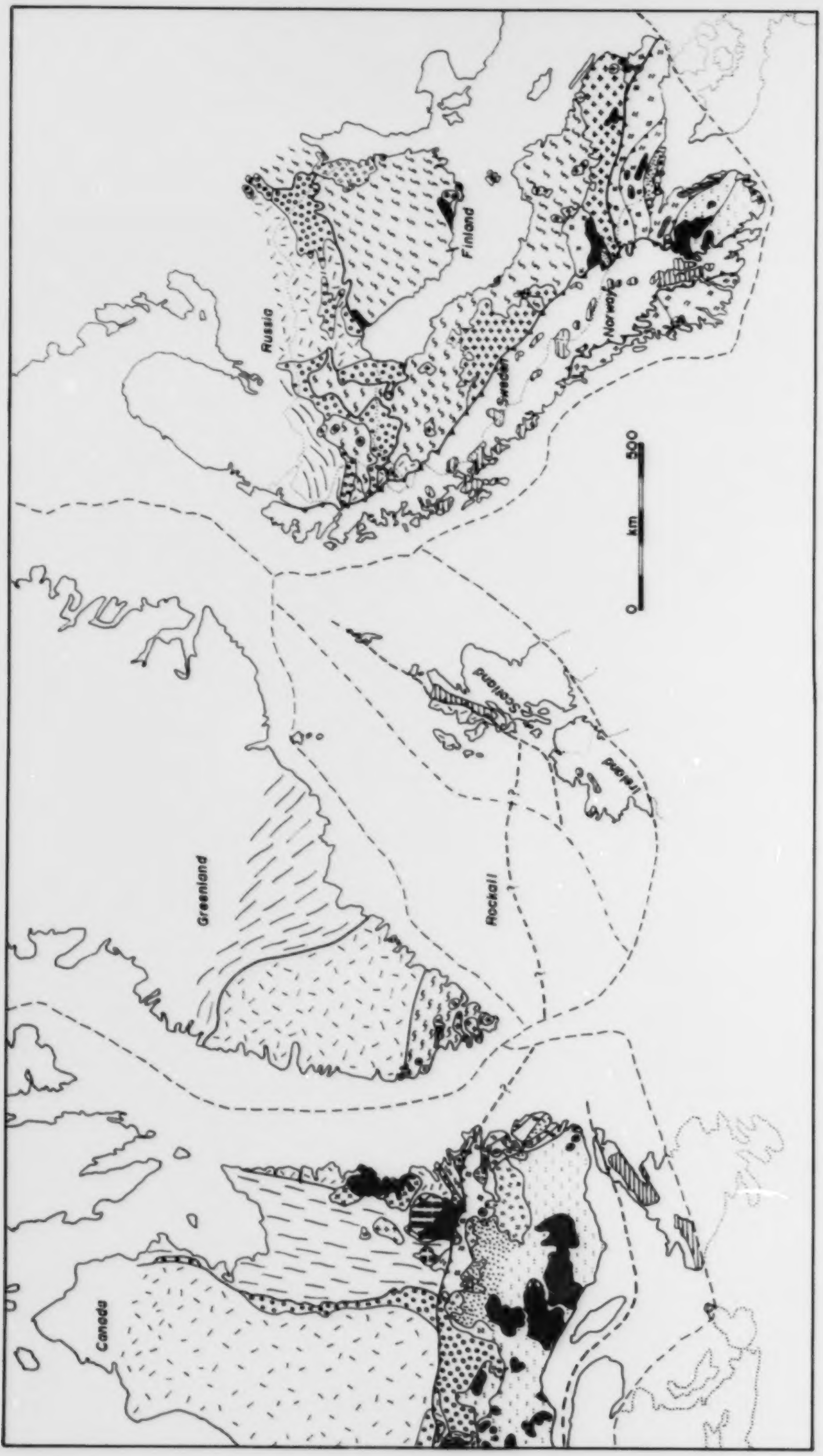
Canada
Greenland
Rockall
Ireland
Scotland
Norway
Sweden
Finland
Russia
0
km
500

Figure 1. *Simplified geological map of Baltica and eastern Laurentia. Modified from Gower et al. (1990).*

Legend (for Figure 1)

Other Units

▨ *Mid-Proterozoic inliers in Caledonian–Appalachian orogen*
░ *Undivided Grenvillian–Sveconorwegian orogen*

Magmatism and orogeny (1.09 to 0.90 Ga)

⁙ *Granitic, syenitic and anorthositic intrusions (Grenvillian and late Sveconorwegian orogenies)*

Extensional tectonism (1.20 to 1.09 Ga)

■ *Anorthosite–mangerite–charnockite–granite (AMCG) suites*
● *Mafic and mafic–felsic supracrustal sequences*

Accretion and orogenesis (1.27 to 1.20 Ga)

Units formed during this period are not distinguished on map. They are within the Grenville–Sveconorwegian orogen

Crustal stability, phase II (1.43 to 1.27 Ga)

■ *Rift-related sedimentation and bimodal volcanism*
◩ *Granite–rhyolite suite / anorthositic–gabbroic intrusions*

Crustal stability, phase I (1.51 to 1.43 Ga)

◩ *Granite plutons / anorthositic–gabbroic intrusions*

Crustal stabilization (1.59 to 1.51 Ga)

* * * *Later rapakivi granite massifs (including Åland, Vehmaa and Laitila massifs)*
⁓ *Granitoid rocks (e.g., southwest Sweden)*

Accretionary and non-accretionary events (1.71 to 1.59 Ga)

░ *Anorthositic, rapakivi granite and gabbronorite massifs (including Wiborg and Mealy Mountains massifs)*
≈ ≈ ≈ *Accretionary volcanic and plutonic rocks (Labradorian, Gothian–Kongsbergian orogenies)*

Accretionary, post-orogenic/anorogenic and rifting environments (1.81 to 1.71 Ga)

░ *Sedimentary sequences (continental margin / rift-related)*
⁙ *Late- and post-Svecofennian/Makkovikian/Ketilidian rocks*

Magmatic arc accretion (1.90 to 1.81 Ga)

▨ *Early Proterozoic arcs south of Archean cratonic nuclei (Svecofennian, Makkovikian, Ketilidian)*
≋ *Reworked Archean and Early Proterozoic crust north of, and between Archean cratonic nuclei*

Pre- 1.90 Ga rocks

⁙ *Continental margin supracrustal sequences*
░ *Archean cratonic nuclei*

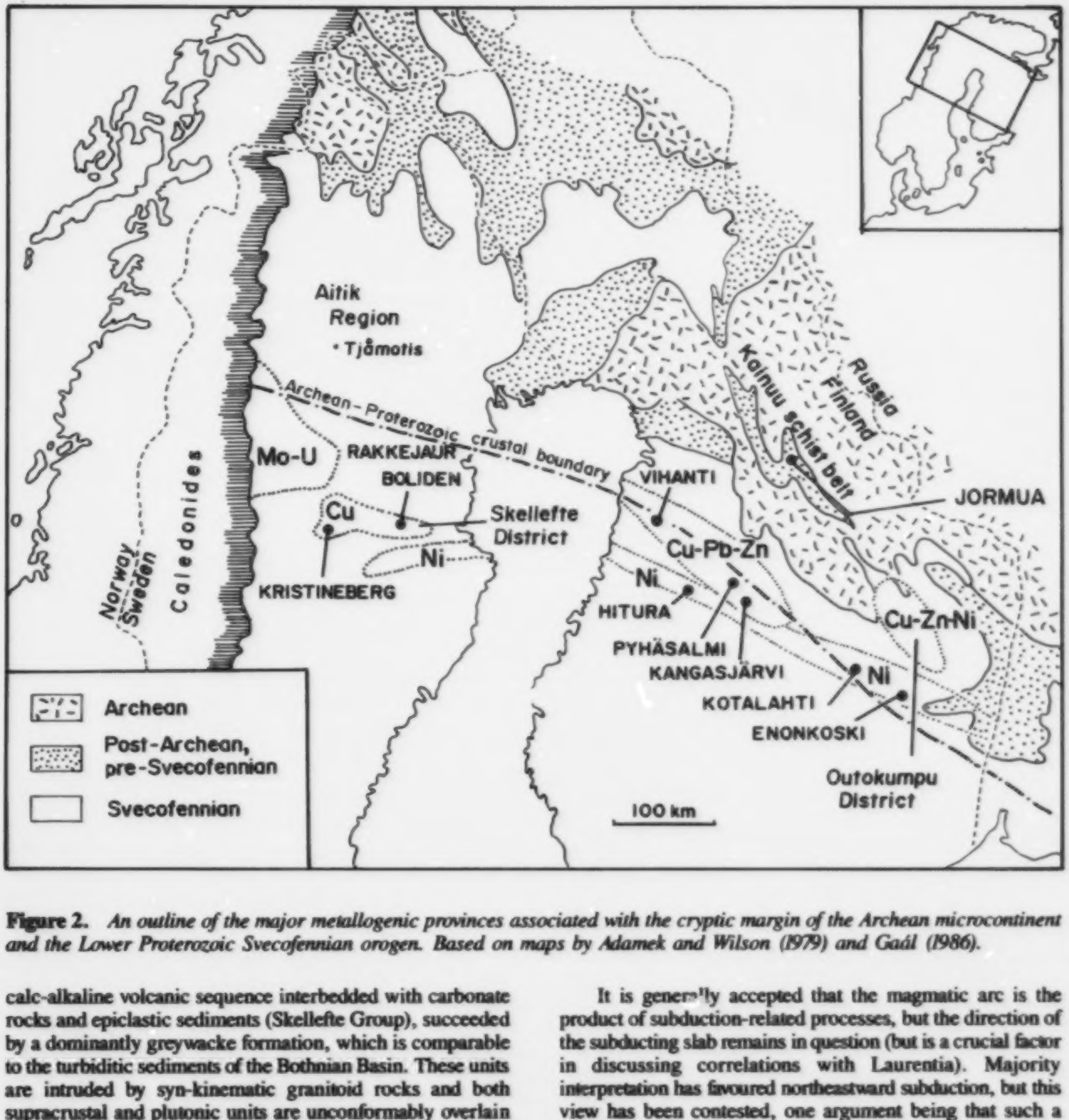
Aitik
Region
Tjåmotis
Archean - Proterozoic crustal boundary
Mo-U
RAKKEJAUR
BOLIDEN
Cu
Skellefte
District
Ni
KRISTINEBERG
Norway
Sweden
Caledonides
Kainuu schist belt
Russia
Finland
JORMUA
VIHANTI
Cu-Pb-Zn
Ni
HITURA
PYHÄSALMI
KANGASJÄRVI
Ni
KOTALAHTI
ENONKOSKI
Cu-Zn-Ni
Outokumpu
District
100 km
Archean
Post-Archean,
pre-Svecofennian
Svecofennian

Figure 2. *An outline of the major metallogenic provinces associated with the cryptic margin of the Archean microcontinent and the Lower Proterozoic Svecofennian orogen. Based on maps by Adamek and Wilson (1979) and Gaál (1986).*

calc-alkaline volcanic sequence interbedded with carbonate rocks and epiclastic sediments (Skellefte Group), succeeded by a dominantly greywacke formation, which is comparable to the turbiditic sediments of the Bothnian Basin. These units are intruded by syn-kinematic granitoid rocks and both supracrustal and plutonic units are unconformably overlain by the dominantly fluvial Vargfors Group. Granitoid rock emplacement postdating the time of arc formation occurred at 1.84 Ga (e.g., Avaviken granite; Wilson *et al.*, 1985) and during late- to post-Sveconfennian events (discussed later). The volcanic rocks in the lower part of the Skellefte Group have island-arc affinities, whereas the upper part of the Skellefte Group has a bimodal rift-related character (Vivallo, 1987; Weihed *et al.*, *in press*). The volcanic arc is succeeded southward by turbiditic sediments of the Bothnian Basin, accompanied by subordinate 1.87 Ga (Welin, 1987) volcanic rocks. In the centre of the basin, the sediments achieve a thickness of 10,000 m (Lundqvist, 1987). Correlative island-arc volcanic rocks in Finland are reviewed by Gaál (1986).

It is generally accepted that the magmatic arc is the product of subduction-related processes, but the direction of the subducting slab remains in question (but is a crucial factor in discussing correlations with Laurentia). Majority interpretation has favoured northeastward subduction, but this view has been contested, one argument being that such a model would not provide any reason for closure of an oceanic environment between the arc and the proto-continent to which the arc was later accreted (Weihed *et al.*, *in press*). Southwestward subduction would allow the Ni-mineralized intrusions (see below) to be placed in a back-arc spreading context. Recent seismic results from the Bothnian sea have indicated a series of deep northeastward-dipping conductors that have been taken as indicating remnant oceanic crust and support for subduction having been northeastward (BABEL Working Group, 1990).

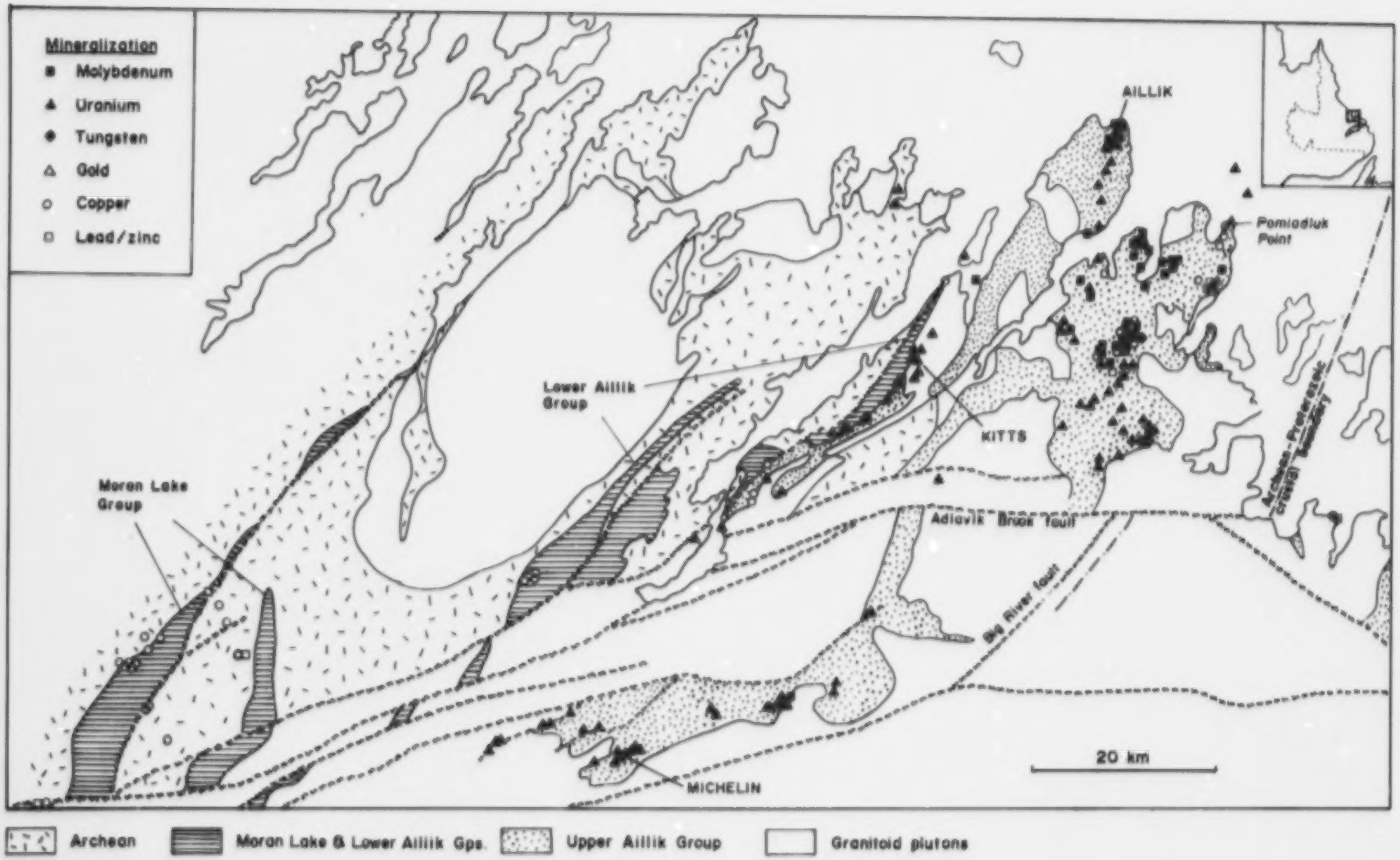
Mineralization
■ Molybdenum
▲ Uranium
◆ Tungsten
△ Gold
○ Copper
□ Lead/zinc
AILLIK
Pomiadluk Point
Archean–Proterozoic crustal boundary
KITTS
Lower Aillik Group
Moran Lake Group
Adlavik Brook fault
Big River fault
MICHELIN
20 km
Archean
Moran Lake & Lower Aillik Gps.
Upper Aillik Group
Granitoid plutons

Figure 3. *Western and central parts of the Makkovik Province, Labrador. Simplified from Gower* et al. *(1982a) and Ryan (1984).*

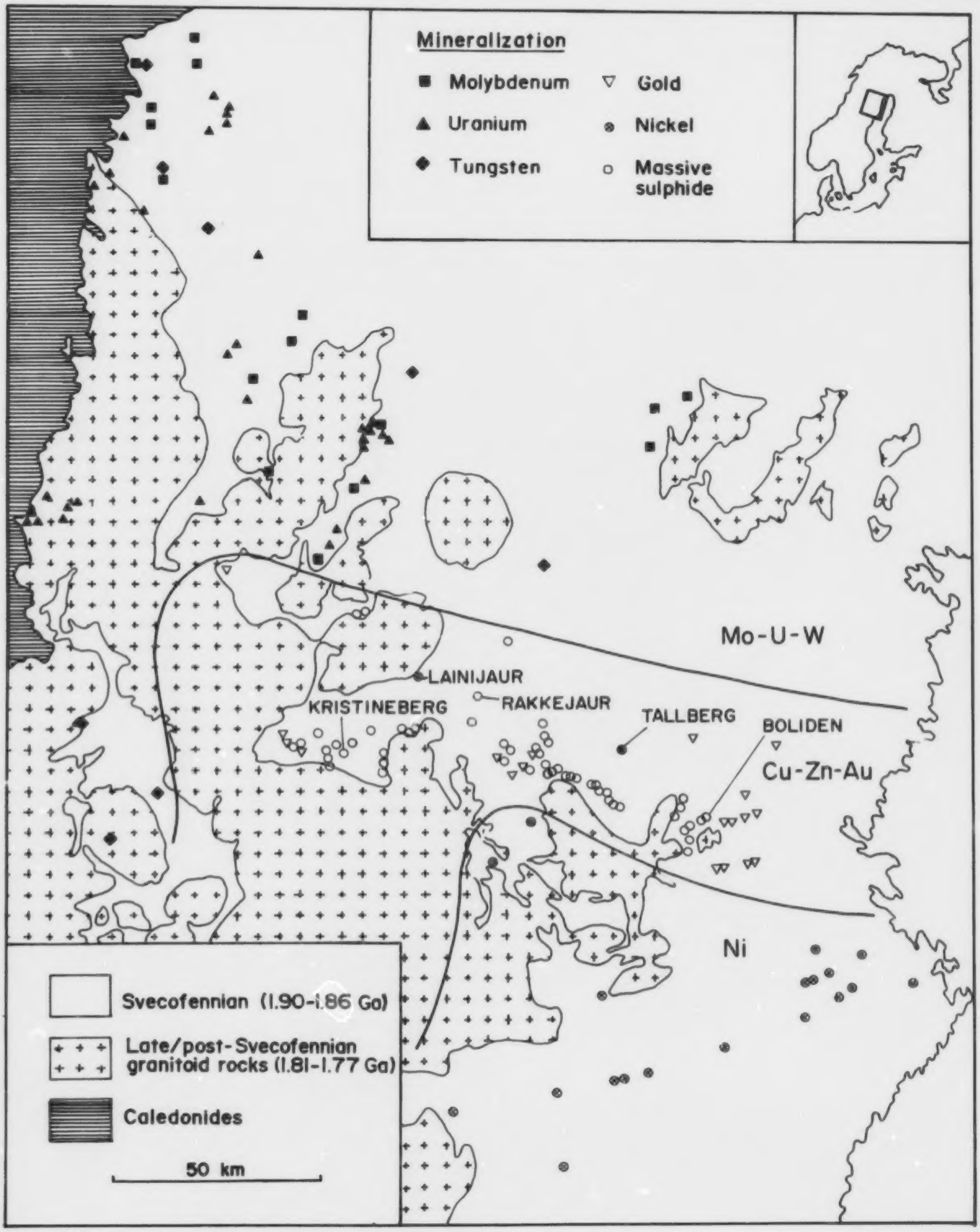
Mineralization
Molybdenum
Uranium
Tungsten
Gold
Nickel
Massive sulphide
Mo-U-W
LAINIJAUR
RAKKEJAUR
KRISTINEBERG
TALLBERG
BOLIDEN
Cu-Zn-Au
Ni
Svecofennian (1.90-1.86 Ga)
Late/post-Svecofennian granitoid rocks (1.81-1.77 Ga)
Caledonides
50 km

Figure 4. *Distribution of mineralization in the Skellefte district. Based on data from Adamek and Wilson (1979), Walser and Einarsson (1982), Öhlander (1986), Rickard (1986), and Weihed* et al. (in press). *The lines separating metallogenic provinces are for visual emphasis only.*

METALLOGENY

Mineralization in the northern Svecofennian arc can be subdivided into four groups, as follows:

(i) Cu–Zn–Au massive sulphide deposits associated with the northern Svecofennian magmatic arc,

(ii) Cu–Mo–(Au) porphyry-type mineralization in felsic intrusions, syn-kinematic with arc volcanism,

(iii) Mo–W mineralization temporally related to Svecofennian orogenesis, but spatially restricted to the Karelian hinterland, and

(iv) Ni mineralization south of the magmatic arc.

Cu–Zn–Au Volcanogenic Massive Sulphides

The Cu–Zn–Au deposits in the northern magmatic arc include those of the Skellefte district and in possibly tectonically correlative regions in Finland (Figure 2). The Skellefte district is currently the most important ore-producing region in Sweden, supporting 10 mines in volcanogenic massive sulphide deposits and two Au open-pit operations (Weihed *et al.*, *in press*). The major deposits in Finland are at Vihanti, Pyhäsalmi and Kangasjärvi (Gaál, 1986). The largest of these orebodies exceeds 20 million tonnes (Kristineberg and Rakkejaur (Skellefte district), and Pyhäsalmi), but most are less than 10 million tonnes, and several less than 1 million tonnes.

In the Skellefte district (Figure 4), the deposits are mostly concentrated at the top of the bimodal upper part of the Skellefte Group. They are associated with felsic pyroclastic rocks, quartz porphyries and minor mafic volcanic rocks in the footwall, and calcareous or graphitic reworked volcanoclastic rocks in the hanging wall. The deposits are typically stratabound and massive, but stringer and disseminated ores are also found. Chalcopyrite and sphalerite are the main ore minerals, and are associated with pyrite (the dominant sulphide), arsenopyrite, galena and pyrrhotite. Gold is an important component of the Skellefte ores and silver is a significant by-product from many deposits (Rickard, 1986; Weihed *et al.*, *in press*). The Skellefte ore bodies are widely interpreted to be volcanogenic massive sulphide deposits (Weihed *et al.*, *in press;* Rickard, 1986; Vivallo, 1987). Rickard (*op. cit.*) envisages a deep marine, hydrothermally dominated process with accumulation of the ore on the seafloor close to volcanic vents.

The deposits in Finland have some similar characteristics, including an association with bimodal volcanic rocks and a Au-rich signature (Gaál, 1986); however, there are differences. The Skellefte district is situated southwest of the cryptic Archean craton margin, whereas the Finnish deposits are situated at the Archean–Proterozoic crustal boundary. Also, the stratigraphic sequence of the Skellefte district is not easily compared with that of the Vihanti-Pyhäsalmi deposits. The latter are better correlated with the Tjåmotis area (Figure 2), 180 km north of the Skellefte district (K. Sundblad, personal communication, 1992). Finally, the Pb isotopic signatures of the Skellefte and Vihanti–Pyhäsalmi districts differ (Sundblad, 1991a).

Cu–Mo–(Au) Mineralization

The Cu–Mo–(Au) mineralization, interpreted to be of porphyry type, is best documented from the Tallberg area (Figure 4) in the 1.89 Ga Jörn batholith (Wilson *et al.*, 1987) in the Skellefte district (Weihed *et al.*, 1987). Other, similar occurrences in the Skellefte district have been interpreted to be porphyry type (Walser and Einarsson, 1982), and comparable mineralization has been reported in Finland (Nurmi and Haapala, 1986). The Tallberg mineralization, first discovered in 1979, has not been exploited, although a large tonnage (45 million tonnes; Weihed *et al.*, *in press*) has been outlined.

The mineralization is hosted by subvolcanic quartz-feldspar stocks (dated at 1886 ±9 Ma; Weihed and Schöberg, 1991) belonging to the oldest phase of the intrusion, and is accompanied by strong propylitic and phyllic alteration. The minerals present include chalcopyrite, pyrite, molybdenite, sphalerite and magnetite concentrated in veins, with or without quartz, calcite and chlorite. The mineralization is low grade (0.27 percent Cu) and of disseminated stockwork type (Claesson and Weihed, 1986; Weihed *et al.*, *in press*). Auriferous quartz ± tourmaline veins are closely associated with the base-metal mineralization. Similar veins are also known in other 1.89 Ga granitoid intrusions (Gaál and Sundblad, 1990). The relatively high Au and low Mo, together with the primitive I-type compositional characteristics of the host rocks, are indicative of an island-arc setting for the Skellefte porphyry mineralization (Weihed *et al.*, *in press*).

Possibly broadly correlative to the porphyry occurrences, are Cu–Au deposits in the Aitik region (Figure 2), hosted by 1.91 Ga sedimentary and felsic volcanic rocks. The deposits are large and low grade, containing disseminated chalcopyrite, gold, bornite, chalcocite, sphalerite, and bismuth- and silver-tellurides (Gaál and Sundblad, 1990).

Mo–(W) Mineralization

Mo–(W) mineralization is widely scattered throughout the Karelian continental domain north of the Skellefte district. Most occurrences are minor and there is no current production of Mo, although some attempts have been made to mine it in the past. The main host rocks are aplite and pegmatite, but mineralization also occurs as impregnations and veins within granite, or in faults transecting altered and metamorphosed volcanic rocks. In addition to molybdenite, some occurrences also have associated chalcopyrite, pyrite, scheelite and fluorite (Öhlander, 1985, 1986). The host granitoid intrusions display within-plate geochemical characteristics, but this does not mean that they are tectonically divorced from the magmatic arc. As Weihed *et al. (in press)* point out, it is common to find such granites on the continental side of arc systems.

Ni Mineralization

Ni mineralization is mostly situated south of the Skellefte district, although a few occurrences are located within it. Many of the occurrences have been known for several decades, but only a few, including the Lainijaur deposit in Sweden (Figure 4) and the Hitura, Kotalahti and Enonkoski deposits in Finland (Figure 2), have progressed to the production stage. The mineralization is associated with ultramafic and mafic rocks intruded into migmatized greywackes and pelites of turbidite affinity. The ultramafic rocks include peridotite, pyroxenite and picrite and the mafic rocks comprise gabbro and norite. Compositionally these rocks range from komatiitic to tholeiitic, and have MORB to island-arc petrotectonic affinity. Massive, vein, breccia and disseminated mineralization is found, containing pentlandite, pyrrhotite and chalcopyrite. The deposits are the product of fractionation of synorogenic tholeiitic intrusions. Gaál (1986) has noted a spatial linkage between the Finnish deposits and dextral transcurrent faults at the edge of the Karelian continent.

LAURENTIAN ANALOGUES (1.91 to 1.84 Ga)

The central part of the Makkovik Province (Figure 3) appears to the closest Laurentian analogue of the Svecofennian northern magmatic arc. Large areas of this region are underlain by the Upper Aillik Group, which consists of felsic volcanic and volcanoclastic rocks associated with cogenetic high-level intrusions and minor mafic volcanic rocks. The contact between the Lower and Upper Aillik groups is highly tectonized, obscuring original (unconformable?) relationships. Dating of the Upper Aillik Group has demonstrated that there were at least two pulses of volcanism, at 1.86 and 1.81 Ga (Schärer *et al.*, 1988). The 1.81 Ga felsic volcanic rocks are the extrusive equivalents of 1.81 to 1.80 Ga granitoid rocks, which are discussed in the next section. Similar dates from granites have been reported from Greenland by van Breemen *et al.* (1974), who obtained ages of 1,845 ± 15 Ma and 1,805 ± 25 Ma.

Granitoid plutonic rocks of several ages are spatially associated with the supracrustal sequences. Two pre-1.80-Ga groups of granitoid rock have been identified in the Makkovik Province and have ages of ca. 1.90 and 1.84 Ga (Kerr *et al.*, *in press*). The latter age group is defined by one date from an intrusion in the eastern part of the Makkovik Province. In terms of correlations between the Makkovik Province and the Svecofennian orogen, the 1.86 Ga age from felsic volcanic rocks of the Upper Aillik Group and the 1.90 and 1.84 Ga ages from granitoid rocks indicate that comparable activity was taking place in both regions simultaneously.

The metallogenic picture for the Upper Aillik Group and associated granitoid rocks is not yet clear. In terms of searching for analogues of the metallogenic styles described above, however, it would seem that the Upper Aillik Group provides a possible analogue for the Skellefte Group, and should be prospected for comparable Cu–Zn–(Au) massive sulphide mineralization. It should be emphasized that, in drawing analogues with the Skellefte Group, only rocks having 1.86 Ga ages should be considered. In the Makkovik Province, these rocks have been only partially distinguished from those having 1.81 Ga ages. Another point is that the Upper Aillik Group appears to be situated above Archean crust, whereas the Skellefte magmatic arc is situated southwest of Archean crust. Given that the Finnish deposits are situated at the Archean–Proterozoic crustal boundary, however, they provide an alternative analogue.

The Cu–Pb–(Au) occurrences that are spatially associated with the Upper Aillik Group, and which might be comparable to either the Skellefte or Vihanti–Pyhäsalmi districts, are situated in easterly areas of the Makkovik district (i.e., south of Pomiadluk Point, Figure 3). Although some of these occurrences are demonstrably post-Makkovikian, their distribution could still be influenced by a Makkovikian or older basement control. It may be significant that they are close to the concealed margin of Archean crust as defined from Nd–Sm and Pb–Pb isotopic evidence (Kerr and Fryer, 1990; Wilton, 1991). Does this mean that Skellefte-type massive sulphides should be sought farther east in the Makkovik Province? If it does, then this is indeed unfortunate because most of the eastern Makkovik Province is underlain by younger granitoid rocks.

On the other hand, a hypothetical location for such a belt allows positioning of other northern Svecofennian metallogenic elements in a Makkovikian context. The Mo–W occurrences that are situated in the Karelian continental domain would equate with Mo occurrences that are mostly present west of a north–south line through Pomiadluk Point, and any search for mafic/ultramafic-hosted Ni deposits would have to focus on the Archean–Proterozoic cryptic crustal boundary. Note that this boundary is displaced dextrally by the Adlavik Brook fault and the location of the Big River fault may be controlled by the margin of the Archean craton. Scope for 1.89 Ga Cu–Mo–(Au) porphyry-type mineralization would seem to be restricted to granitoid rocks in the southwest and western parts of the Makkovik Province on the basis of present geochronological data.

LATE/POST- SVECOFENNIAN– LATE/POST- MAKKOVIKIAN

Until recently, very few occurrence of Li and Sn mineralizations were known in Sweden... (Lundqvist *et al.*, 1986).

GEOLOGICAL SETTING

As used in this report, late- and post-Svecofennian rocks refer to those emplaced between 1.81 and 1.77 Ga. Included, therefore, are the granites within and north of the Skellefte district, the 'Revsund-type' granitoid rocks in the central Svecofennian orogen south of the Skellefte district (Lundqvist, 1990), much of the Transscandinavian Igneous Belt, and scattered intrusions in eastern Baltica. The term 'Transscandinavian Igneous Belt' is used in this report for 1.81 to 1.77 Ga granitoid rocks distributed along the western flank of the Svecofennian orogen, extending from southeastern Sweden, northward under the Caledonides, to the Lofoten Islands in Norway (Gorbatschev, 1985; Gaál and Gorbatschev, 1987). As defined in Scandinavia, the Transscandinavian Igneous Belt includes the 1.71 to 1.65 Ga Dala volcanic rocks and associated granitoid intrusions. The present author, however, has chosen to exclude the Dala region from the Transscandinavian Igneous belt (see page 345).

The Revsund intrusions consist mostly of massive granite, but minor granodioritic to quartz-monzonitic rock types are also present (Lundqvist, 1990). The intrusions within and north of the Skellefte district have bulk compositions similar to the 'Revsund-type' granites and have been compared closely with them. The granitoid rocks emplaced through pre-Svecofennian crust north of the Skellefte district commonly have late-stage minor intrusions and mineralization (see below), and are interpreted to have been derived from an underlying Archean crustal source (Öhlander, 1986). Similarly, the 'Revsund-type' granites in the Proterozoic Bothnian Basin are considered to have been derived largely from sedimentary rocks.

The 1.81 to 1.77 Ga granitoid rocks of the Transscandinavian Igneous Belt have, predominantly, granite to quartz-monzonite compositions, but monzodioritic and dioritic phases are present in places (Gaál and Gorbatschev, 1987). Traditionally, the rocks have been divided into textural types—the K-feldspar megacrystic 'Filipstad' type, and the coarse-grained, even-grained 'Växjö' type. The Växjö type is generally closer to granite *sensu stricto*, and is commonly the younger type. The Transscandinavian Igneous Belt granitoid rocks lack pegmatite or other minor intrusions. Volatile deficiency, coupled with compositional characteristics, is interpreted to indicate derivation from deep crustal sources (Lindh and Persson, 1990). A deep crustal source interpretation is consistent with other features, such as having structures conformable with their host rocks and showing common evidence of mafic-felsic magma interaction (cf. Andersson, 1991).

METALLOGENY

It should be stressed at the outset that the late- and post-Svecofennian granites in the north are of most interest in a metallogenic context. In contrast, the Transscandinavian Igneous Belt is generally barren. Mineralization in the northern late- and post-Svecofennian 1.81 to 1.77 Ga granitoid rocks is typically granophile. All of the occurrences are fairly small, but some are potentially large enough to be economically viable. The types of mineralization reflect the variation in the source rocks from which the host granites were derived. The following metallogenic groupings can be recognized:

(i) Mo–U–(W)–(Sn) mineralization associated with granitoid rocks emplaced into Karelian crust, and

(ii) W–Sn and Sn–Li mineralization in granitoid rocks in the Bothnian Basin and farther south.

Although the focus here is on granitoid rocks emplaced between 1.81 and 1.77 Ga, it should be noted that not all the mineralization is confined to granites of this age. Öhlander and Billström (1989) have investigated two granites, both of which host Mo-mineralized aplites, and found them to have ages of 1.89 and 1.77 Ga, respectively. The Mo-mineralized aplite in the 1.89 Ga granite grades into its host, and hence cannot be a later dyke. Despite this example, Öhlander and Billström (1989) concluded that most of the Mo- and W-occurrences are associated with the younger 1.81 to 1.77 Ga granites. This judgement is supported by Wilson and Åkerblom (1982), who summarized data demonstrating that the age of associated U mineralization was ca. 1.75 Ga.

Mo–U–(W)–(Sn) Mineralization

Difficulties in discriminating between the 1.89 and 1.86 Ga and 1.81 to 1.77 Ga Mo–W occurrences are a consequence of the morphological similarities between the two age groups, hence the reader is referred back to the previous section for descriptive details. Information regarding the U mineralization is given here, however – heeding the geochronological data for its younger age.

The major Mo deposits (Figure 5) include Björntjärn (1 million tonnes at ca. 0.15 to 20 percent Mo) and Munka (1.5 million tonnes at 0.10 to 0.15 percent Mo), both of which have been genetically related to the younger granites (Walser and Einarsson, 1982).

The most important U mineralization (Figure 5) includes the Pleutajokk deposit (6,000 tonnes U) and the Björklund and Rävaberget occurrences (Öhlander, 1986). The mineralization occurs as epigenetic fracture fillings and in disseminated form within Na-metasomatized rhyolitic volcanic rocks or granites. The generally thin veins occur singly and in stockworks. Uraninite, the ore mineral, is associated with calcite, fluorite, hematite and uranotitanite. The W occurrences are spatially related to the U and Mo mineralization.

Although geochronological linkage to the younger granites is recognized, most workers consider that the mineralization is the result of a multi-stage process. This process involved scavenging of granophile elements from pre-Svecofennian sialic crust into the 1.89 to 1.86 Ga granitoid rocks and then mobilization of the same elements during later 1.81 to 1.77 Ga granitoid activity (Öhlander, 1986). Thus, from the prospecting point of view, although the earlier stage of scavenging could lead directly to concentrations of economic interest, the younger intrusions are the better targets.

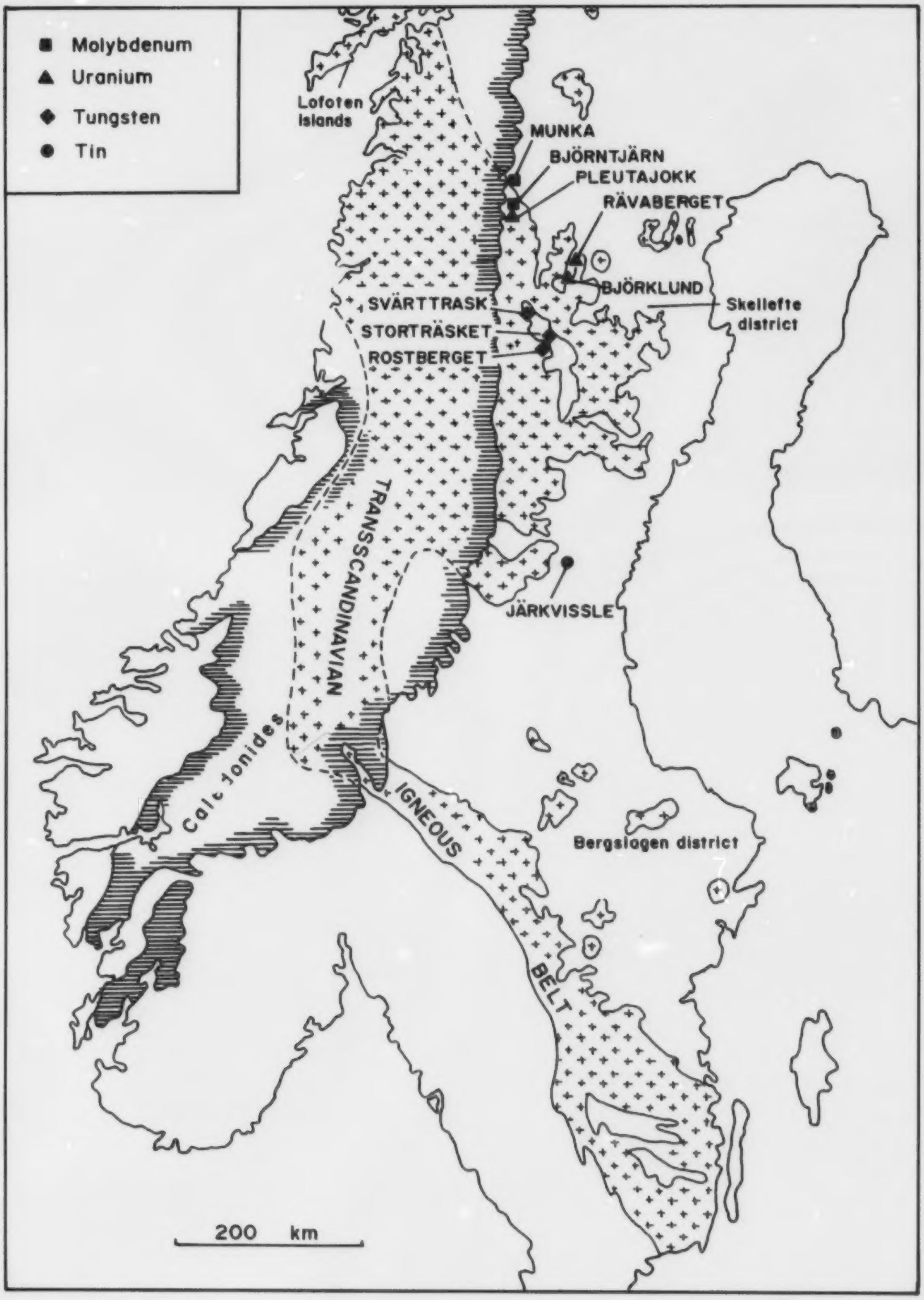
■ Molybdenum
▲ Uranium
◆ Tungsten
● Tin
Lofoten Islands
MUNKA
BJÖRNTJÄRN
PLEUTAJOKK
RÄVABERGET
BJÖRKLUND
Skellefte district
SVÄRTTRASK
STORTRÄSKET
ROSTBERGET
TRANSSCANDINAVIAN
JÄRKVISSLE
Caledonides
IGNEOUS
BELT
Bergslagen district
200 km

Figure 5. *Distribution of late- and post-Svecofennian rocks in the Baltic Shield and associated major mineral deposits.*

W–(Sn) and Sn–Li Mineralization

The association of W–(Sn) mineralization with 1.81 to 1.77 Ga granitoid rocks is well established. Outside the Bergslagen district, the best described areas of mineralization are south of the Skellefte district (Öhlander, 1986). Some of the deposits have been explored in detail, including extensive diamond drilling, but none have been developed.

The typical mode of occurrence is greisen and quartz veins in granite (Rostberget, Storträsket; Figure 5) and in country rocks close to granite. Wolframite is typically the most important W ore mineral, but scheelite dominates at Svärtträsk. Associated minerals in some occurrences include arsenopyrite, molybdenite, pyrite, pyrrhotite, chalcopyrite and galena. Fluorite, allanite, epidote, apatite and ilmenite are common accessory minerals and tourmaline, xenotime and topaz have also been reported.

Another closely allied metallogenic association is Sn–Li mineralization, which has been recently discovered in 1.82 Ga (late-Svecofennian) granitoid rocks in central Sweden (Lundqvist *et al.*, 1986). Some of the mineralization is potentially of economic interest (e.g., Järkvissle; Figure 5), although no development work has been carried out. The mineralization occurs in muscovite pegmatites intruded into metagreywacke in the Bothnian Basin. The economically interesting minerals include spodumene, petalite, cassiterite and columbite–tantalite.

LAURENTIAN ANALOGUES (1.81 to 1.77 Ga)

The Makkovik Province (Figure 3) offers the closest tectonic analogue for the late-Svecofennian and Transscandinavian 1.81 to 1.77 Ga rocks. Granitoid rocks having ages of 1.80 Ga are now known to be widespread in the central Makkovik Province (Kerr and Krogh, 1990; Kerr *et al.*, *in press*), and a sub-volcanic quartz–feldspar porphyry of the Upper Aillik Group has yielded a similar age (1.81 Ga) (Schärer *et al.*, 1988). Like their late-Svecofennian counterparts, these granitoid rocks are generally undeformed and have dominantly granite compositions, associated with lesser syenite, monzonite, quartz monzonite and monzogranite. The rocks have transitional calc-alkaline to alkaline character.

The Makkovik area is well known for its Mo–U mineralization, with which less well-known W–Cu–Zn–Pb mineralization is also associated (Gower *et al.*, 1982a; MacDougall and Wilton, 1987, 1988; Wilton and Wardle, 1987). The mineralization is hosted by metasedimentary rocks in the Lower Aillik Group (U + pyrite) and by felsic volcanic or volcanoclastic rocks in the Upper Aillik Group. Locally, some occurrences are hosted by pegmatite, granite and amphibolite. The most important deposits are Kitts (207 000 tonnes at 0.73 percent U_2O_8), Michelin (7 million tonnes at 0.13 percent U_2O_8), and Aillik (2 million tonnes at 0.25 percent Mo) (Gower *et al.*, 1982a). The locations of these deposits are shown in Figure 3 and details of other deposits in the region are given by Gower *et al.* (1982a).

Earlier interpretations that envisaged syngenetic mineralization by hydrothermal leaching of volcanic glasses have lost support in favour of models emphasizing the role of fluids associated with later granitoid rocks. The along-strike continuity within a particular supracrustal unit for several of the deposits remains an argument in favour of the syngenetic model, but metallogenic zonation around a high-level granite (MacDougall and Wilton, 1987) demonstrates a linkage with (post-1.80 Ga) plutonism (MacKenzie and Wilton, 1987; Wilton and Wardle, 1987). The role of granitoid rocks received little attention during earlier exploration and, as a result, many targets may have been underrated, or overlooked entirely.

There may be elements of validity to both the syngenetic and epigenetic models. Scavenging of granophile elements from the Archean basement could have led to Mo–U-enriched 1.91 to 1.86 Ga felsic volcanic and granitic rocks, with further mobilization and redistribution during later magmatic events. The key factor is a multistage process. In this regard, the Makkovik area must be regarded very favourably, as there were periods of magmatism at 1.91 to 1.89 Ga, 1.86 Ga, 1.84 Ga, 1.80 Ga, 1.72 Ga and 1.65 Ga (Kerr *et al.*, *in press;* Schärer *et al.*, 1988). Although granitoid magmatism having ages of 1.72 Ga and 1.65 Ga goes beyond the temporal realm discussed here, it is mentioned because it provided concentration opportunities for metal concentration that were not available in areas of Baltica having Archean crust as its foundation. The Makkovik Province, therefore, may well have greater potential than Baltica for this type of deposit.

STORA LE-MARSTRAND FORMATION–PRE-LABRADORIAN SEDIMENTS

The Modum cobalt deposits were discovered in 1772 and mined until 1898... (Bugge, 1978).

GEOLOGICAL SETTING

Supracrustal rocks of the Stora Le-Marstrand Formation and approximately temporally correlative rocks in the Bamble–Kongsberg sectors (Figure 6) were deposited either coeval with, or shortly after, the formation of the Transscandinavian Igneous Belt and related rocks, but prior to the onset of the Gothian–Kongsbergian orogeny (Åhäll *et al.*, 1990).

The Stora Le-Marstrand Formation consists of migmatized greywacke-dominated sedimentary rocks, associated with banded amphibolites derived from pillowed mafic lavas and volcanoclastic rocks. These rocks are intruded by mafic sills, dykes and small ultramafic to gabbroic bodies. The supracrustal rocks in the Bamble–Kongsberg sectors consist of a lower sequence of quartzite and sillimanite-bearing rocks and an upper division of variable gneisses made up of biotite-, graphite- and sulphide-bearing schist, quartzite and quartz-rich gneiss, with calc-silicate and marble layers (Starmer, 1990, 1991). The age of the Stora Le-Marstrand Formation is poorly constrained by a 1758 ± 58 Ma Sm–Nd

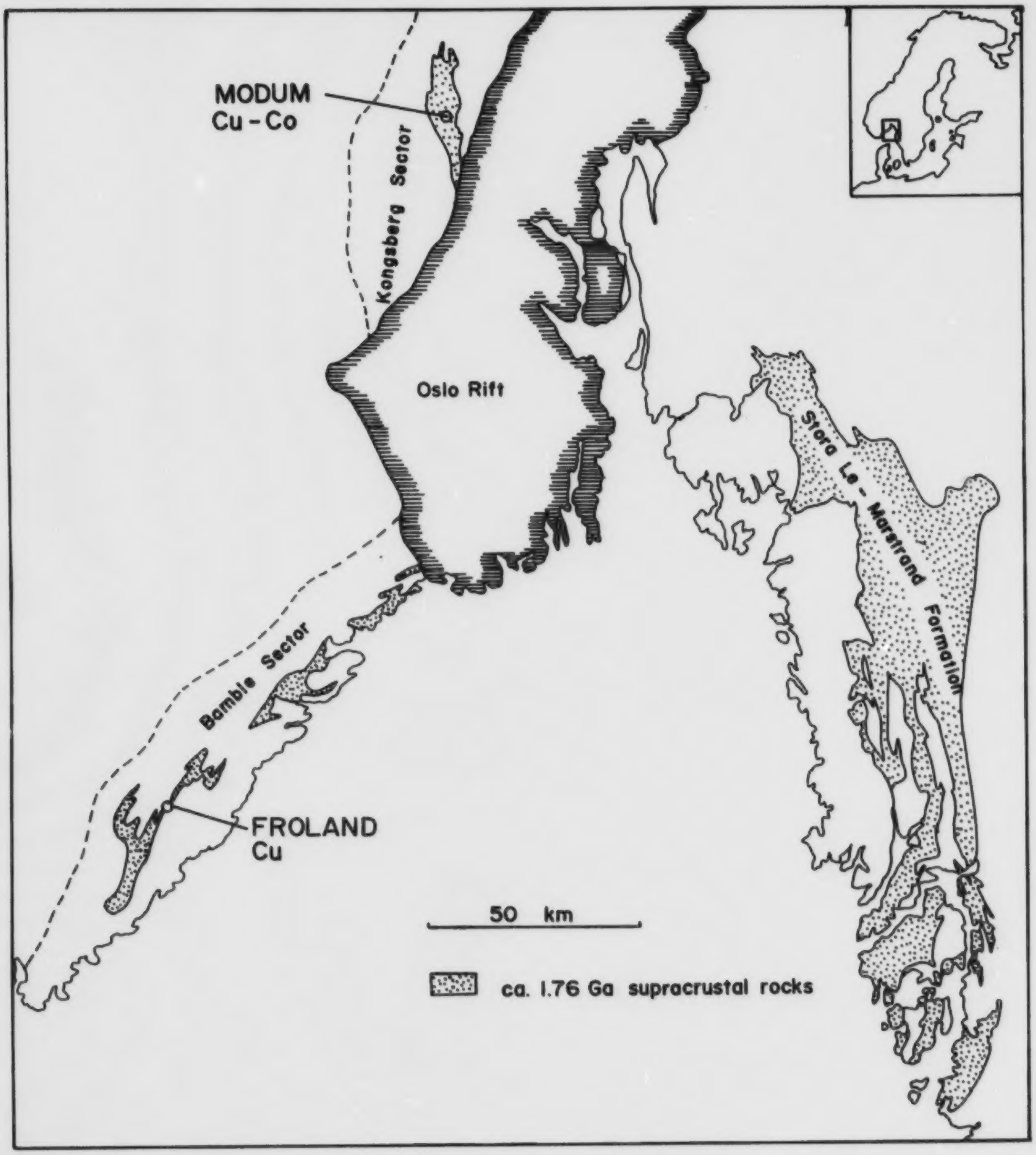
MODUM
Cu - Co
Kongsberg Sector
Oslo Rift
Bamble Sector
FROLAND
Cu
Stora Le - Marstrand Formation
50 km
ca. 1.76 Ga supracrustal rocks

Figure 6. *Distribution of the Stora Le-Marstrand Formation and approximately coeval supracrustal rocks in the Bamble–Kongsberg Sector, and the location of the past-producing Modum and Froland Cu–(Co) deposits.*

errorchron for interstratified mafic volcanic rocks (Åhäll and Daly, 1989). The possibly younger supracrustal rocks in the Bamble–Kongsberg sectors are considered to have been deposited prior to 1680 Ma (Starmer, 1990).

Åhäll and Daly (1989) suggest that the Stora Le-Marstrand Formation had an oceanic-arc or back-arc tectonic setting, but in close proximity to proto-Baltica. The higher proportion of more mature quartz-rich sediments and a dearth of mafic volcanic units in the supracrustal rocks of the Bamble–Kongsberg sectors indicate a somewhat different depositional setting, but one that also involved derivation of material from a nearby sialic source, which is considered by Starmer (1990) to be proto-Baltica.

METALLOGENY

Mineralization linked to the ca. 1760 to 1700 Ma clastic supracrustal rocks in southwest Scandinavia is of one type only, namely stratabound Cu–Co deposits. Deposits located at Modum in the Kongsberg Sector and at Froland in the Bamble Sector were mined in the past. The deposits occur in fahlbands, which are schistose, sulphide-impregnated zones that may extend for several kilometres along strike and be several hundreds of metres thick. The host rocks are sillimanite–cordierite-bearing mica schist, quartzite and amphibolite. The dominant sulphides are pyrite and pyrrhotite, but chalcopyrite, sphalerite, galena, cobaltite, skutterudite, and native bismuth, copper and molybdenite have also been found. The deposits are low grade; the ore at Modum averaged 0.15 percent Cu and 0.1 percent Co. The deposits are believed to have formed from volcanic exhalative or biogenic processes (Bugge, 1978). In current nomenclature, the mineralization could be interpreted as Besshi-type or sedimentary exhalative, depending on the metal associations and extent of volcanic activity.

LAURENTIAN ANALOGUES (1.76 to 1.70 Ga)

The pre- or syn-Labradorian metasedimentary gneisses within the eastern Grenville Province are remarkably similar to their pre-Gothian–Kongsbergian temporal equivalents in Baltica. The Laurentian analogues underlie large tracts of southern Labrador and comprise mainly pelitic to semipelitic gneiss, with minor associated calc-silicate rocks, amphibolite and quartzite. The amphibolite is derived, in part, from mafic volcanic rocks and at least some of the quartzite probably originated as chert (Gower *et al.*, 1987).

In terms of known occurrences of sulphides, the Paradise metasedimentary gneiss belt in southeast Labrador (Figure 7) seems to be one of the best candidates for economic mineralization of the type described above. Spectacular, ochreous-weathering, sulphide-rich (mainly pyrite) zones extending up to 3 km in length are abundant in the Dead Islands and Occasional Harbour areas, and similar occurrences are also known in the vicinity of southern Sandwich Bay. It is worth emphasizing that the number of occurrences correlates directly with the quality of exposure, which deteriorates rapidly away from the coast. The regional continuity of rock types provides a good reason to believe that sulphide-rich zones are present throughout the entire belt (Gower and Swinden, 1991). Although Fe-sulphides are most abundant, traces of Cu are known locally (Gower *et al.*, 1982b, 1987), and Co anomalies are seen in lake-sediment geochemical data. Shallow drilling was carried out at Eagle River (Bradley, 1966), where a surface sample averaged 2.8 percent Cu (Douglas, 1953).

Thought should be given to the possibility that two separate sedimentary environments, possibly having different ages, are present in both the Gothian–Kongsbergian and Labrador orogens. Perhaps the supracrustal rocks in both the Bamble and Kongsberg sectors and the Paradise River metasedimentary gneiss were deposited in rift-related settings (the zones of crustal weakness so formed being subsequently utilized and disguised by later tectonism), whereas the Stora Le-Marstrand Formation and areas outside the Paradise metasedimentary gneiss belt may have had an intra-arc or continental margin setting. Such a distinction could offer an explanation for the presence of sulphide-rich sediments in the Paradise metasedimentary gneiss belt and the Bamble and Kongsberg sectors, but their apparent dearth elsewhere. If there is any validity to this suggestion, it should also be kept in mind that the Paradise metasedimentary gneiss belt is geologically simpler and has a much larger prospective area than the Bamble–Kongsberg sectors.

GOTHIAN–KONGSBERGIAN–LABRADORIAN

GEOLOGICAL SETTING

The Gothian–Kongsbergian orogeny extended from 1.71 to 1.57 Ga and resulted in the accretion of a large area of juvenile crust to the western margin of proto-Baltica (Figure 8). The main area is in southwest Sweden and southeast Norway, but this area probably links up, under the Caledonides, with coeval rocks of the Western Gneiss Region in western Norway. The accreted crust consists dominantly of calc-alkaline plutonic rocks, that, typically, are now orthogneisses. These range from tonalite to granite and include minor anorthositic to gabbroic intrusions. Felsic volcanic rocks and associated sediments are also present (Gaál and Gorbatschev, 1987). Several cycles of plutonism occurred, perhaps at ca. 1.71, 1.66 and 1.60 Ga, but the duration and areal extent of these remains uncertain (cf. Åhäll *et al.*, 1990: Åhäll and Persson, *in press;* Larson *et al.*, 1990, Tucker *et al.*, 1990).

Inboard of the main calc-alkaline belt, partially within the older Svecofennian orogen, is a separate assemblage of felsic volcanic rocks and coeval granitoid intrusions (Figure 8). The felsic extrusive rocks include the Dala porphyries, deposited unconformably on rocks of the Svecofennian orogen (Lundqvist, 1968), and the associated granite intrusions include the Siljan and Garberg plutons. Some of the granites have rapakivi textures (Sundblad, 1991b). The rocks are usually grouped with the Transscandinavian Igneous Belt and, although many Scandinavian workers will accept that they are distinct from the 1.81 to 1.77 Ga parts of the Transscandinavian Igneous Belt (cf. Sundblad, 1991b), not all would agree with grouping them, as has been done here, with rocks of contrasting character west of the Protogine zone.

METALLOGENY

The principal metallogenic feature of the 1.71 to 1.57 Ga Gothian–Kongsbergian orogenic period is an almost complete

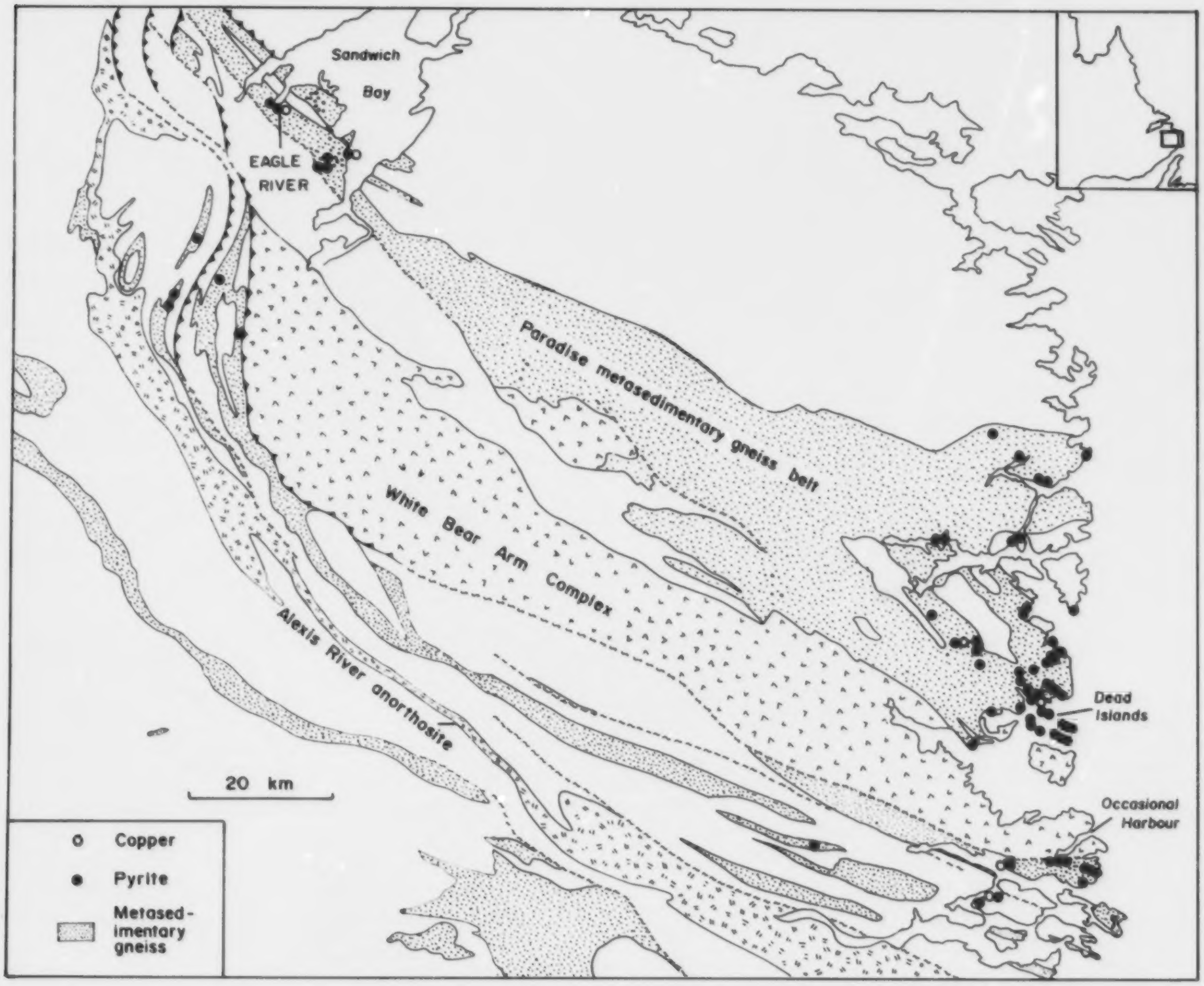
Sandwich Bay
EAGLE RIVER
Paradise metasedimentary gneiss belt
White Bear Arm Complex
Alexis River anorthosite
Dead Islands
Occasional Harbour
20 km
○ Copper
● Pyrite
▨ Metasedimentary gneiss

Figure 7. *Paradise metasedimentary gneiss belt and other nearby major units, including the distribution of pyritic gossans.*

lack of contemporaneous mineralization. There are two noteworthy exceptions, both east of the Protogine zone:

(i) Sn-, Pb-, Zn- and Be-bearing greisen veins in Siljan granite (Figure 8), and

(ii) minor Cu–Zn–Pb deposits hosted by the Dala volcanic rocks.

Sn–Pb–Zn–Be Mineralization

Sn–Pb–Zn–Be mineralization in greisen veins has received only brief mention in the Scandinavian literature (Sundblad, 1991b; Ahl, 1991). The occurrences are found in the Siljan granite, which has an age of ca. 1.70 Ga. Sundblad (1991b) has suggested a linkage with similar mineralization found in younger (less than 1.65 Ga) rapakivi granitoid intrusions in Finland.

Cu–Zn–Pb Mineralization

The Cu–Zn–Pb occurrences include Börningsberget, Rotendal and Storharn in the Dala volcanic rocks and Van and Öradtjärn in the closely related Siljan granite (Johansson and Rickard, 1985). The deposits in the Dala volcanic rocks have been locally exploited for their Cu content (Lundqvist, 1968). Most of the mineralization occurs in the matrix of volcanic breccias and in irregular veins, the latter typically quartz-filled and locally brecciated. It consists of chalcopyrite, chalcocite, sphalerite, galena and pyrite, with associated fluorite, carbonate, epidote, feldspar and quartz. The mineralization and accompanying wall-rock alteration, are interpreted as products of hydrothermal activity. The Pb isotopic studies indicate that the metals were derived from both Svecofennian crustal rocks and post-Svecofennian mantle sources (Johansson and Rickard, 1985).

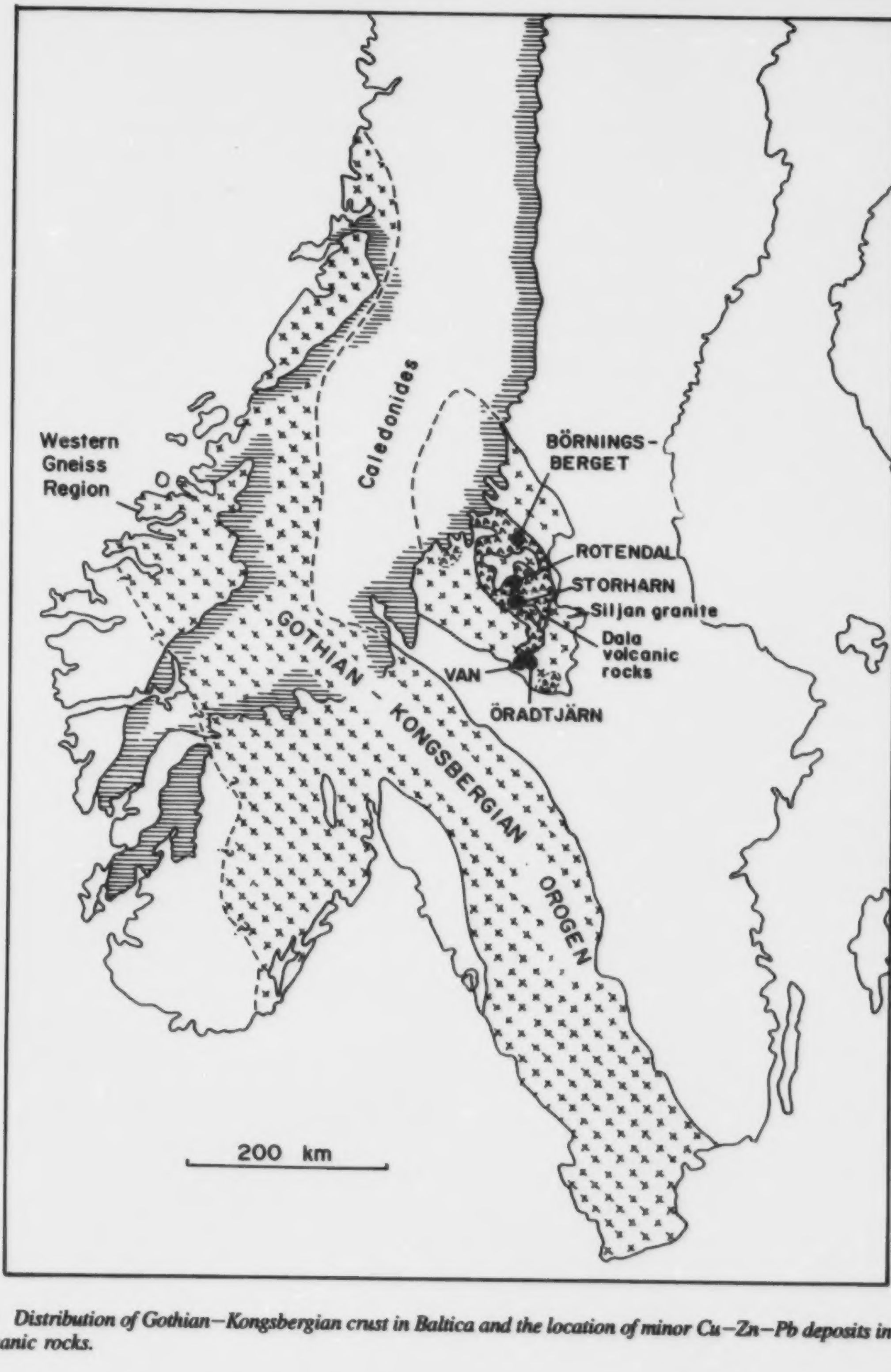
Western
Gneiss
Region
Caledonides
BÖRNINGS-
BERGET
ROTENDAL
STORHARN
Siljan granite
Dala
volcanic
rocks
VAN
ÖRADTJÄRN
GOTHIAN - KONGSBERGIAN OROGEN
200 km

Figure 8. *Distribution of Gothian–Kongsbergian crust in Baltica and the location of minor Cu–Zn–Pb deposits in the coeval Dala volcanic rocks.*

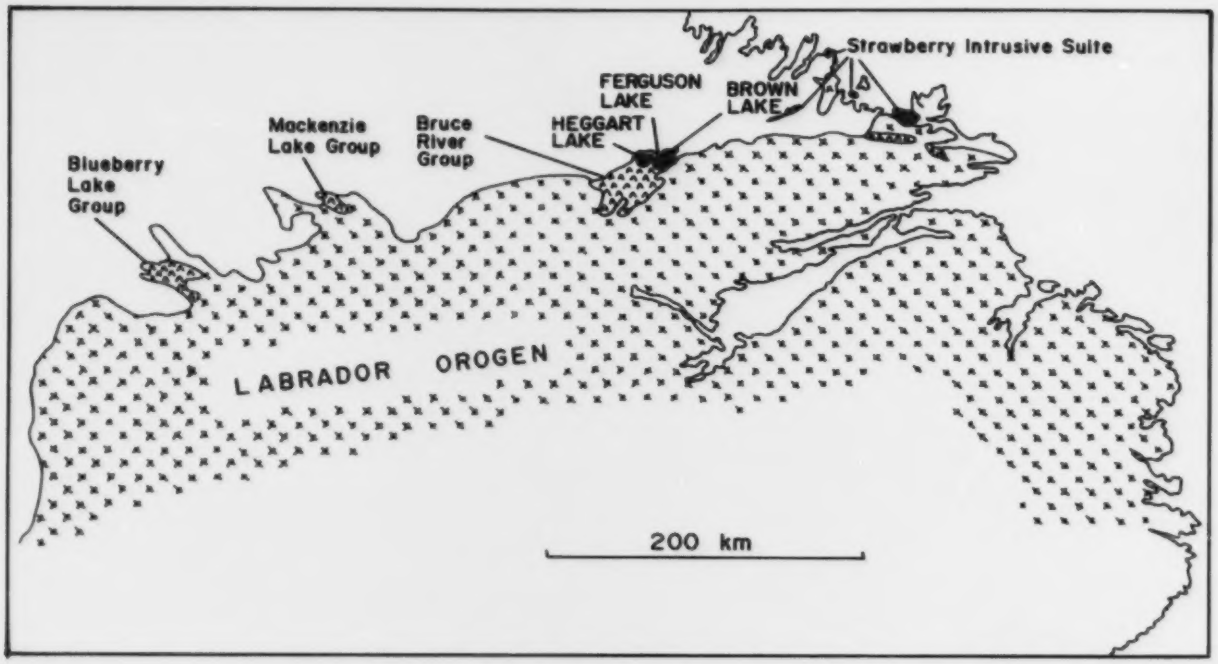

Figure 9. *The distribution of Labradorian crust in Labrador. The locations of the Strawberry Intrusive Suite and felsic volcanic rocks at the northern margin of the Trans-Labrador batholith are also shown, and major mineral occurrences in the Bruce River Group indicated.*

LAURENTIAN ANALOGUES (1.71 to 1.62 Ga)

Rocks formed during the 1.71 to 1.62 Ga Labradorian orogeny (Figure 9) are a close analogue to those of the Gothian–Kongsbergian orogenic belt. Like their Scandinavian plutonic counterparts, they, too, seem to be remarkably devoid of mineralization. Based on the proposed correlations, mineral exploration of Labradorian-age plutonic rocks does not appear promising (this conclusion does not extend to post-Labradorian rocks in the Labrador orogen, however; see below).

An explanation for the dearth of mineralization in rocks of these orogenic belts is not difficult to find, noting that scavenging from older into younger crustal rocks seems to have been an important mechanism in the control of earlier granophile-type mineralization, as discussed earlier. In both Baltica (Lindh and Persson, 1990) and Laurentia (Schärer, 1991), the rocks have been interpreted as juvenile. The only opportunities for a scavenging mechanism to have been operative would have been at the margin of the 1.71 to 1.62 Ga orogenic belts, where magmas passed through older crust. In Scandinavia, the mineralized Dala volcanic rocks and associated granitoid intrusions occupy such a tectonic setting.

In Labrador, two groups of rocks occur in a similar setting and have potential economic significance. These are (i) the Strawberry Intrusive Suite and related rocks in the Makkovik Province and (ii) felsic volcanic rocks of the Bruce River, Mackenzie Lake and Blueberry groups, on the foreland side of the Trans-Labrador batholith.

Like the Siljan granite, the tectonic status of the Strawberry Intrusive Suite is still under discussion. Linkage with Labradorian orogenesis is not necessarily implied here, despite including it in this section. It should be kept in mind, however, that the earliest rocks belonging to the Labrador orogen (1.71 Ga) are not much younger. The 1.72 Ga Strawberry Intrusive Suite has been identified as part of a group of specialized granites and potential for Mo, Cu, Pb and Zn mineralization has already been established (Kerr, 1988). The Strawberry Intrusive Suite is not known to have Sn mineralization, but the Cape Strawberry granite does have anomalously high background Sn concentrations.

Of the felsic volcanic rocks on the foreland side of the Trans-Labrador batholith, only in the Bruce River Group has there been any serious prospecting. This resulted in the discovery of numerous Cu–Zn–Pb–U showings–see summaries by Ryan (1984), Ryan *et al.* (1987) and Wilton (1988). The most significant showings are at Heggart Lake, Ferguson Lake and Brown Lake (Figure 9). The mineralization has been modelled by Wilton (1988) as epigenetic and the product of Grenvillian tectonism. The role of older source rocks and processes should not be dismissed, however, especially as similar mineralization is not ubiquitous throughout the Grenville orogen.

1.65 to 1.27 Ga ANOROGENIC SUITES

> *... in general, the rapakivi granites have been regarded as the most barren rocks of Finland. However, the exploration work carried out ... since 1967 has shown that tin and related beryllium, tungsten, zinc and copper mineralizations often occur in ... the rapakivi massifs.* (Haapala, 1977).

GEOLOGICAL SETTING

1.65 to 1.51 Ga

Contemporaneously with calc-alkaline plutonism at the western margin of Baltica, magmatism of an anorogenic anorthosite–mangerite–charnockite–granite (AMCG) character was taking place within it, about 700 km farther east (Figure 10). The products of this plutonism included the Wiborg rapakivi massif and associated satellite bodies surrounding it (Bodom, Ahvenisto, Suomenniemi, Obbnäs, Onas). The Wiborg massif was emplaced between 1.65 and 1.63 Ga, but some porphyries were emplaced as late as 1.615 Ga (Vaasjoki *et al.*, 1991). Classic rapakivi granite makes up about 80 percent of the body, the remainder being a variety of granite types. This range of rock types is also seen in the satellite intrusions, together with gabbroic, anorthositic and monzodioritic rocks. Remnants of a volcanic carapace to the Wiborg massif are preserved near its southern margin (cf. Rämö and Haapala, 1990). Associated with the Wiborg massif are the northwest-trending Häme diabase and quartz porphyry dykes, dated at 1.64 Ga.

An essentially compositionally identical group of rocks were emplaced farther west about 50 million years later. Included here are the large Åland, Vehmaa, Laitila massifs and the smaller Kökar, Fjälskär, Eurajoki, Kokemäki and Siipyy intrusions in Finland. The Nordsjö, Ragunda, Nordingrå, Rödön and Gävle bodies in Sweden are provisionally included here, although only the Nordingrå body has been dated so far (Figure 10). The age range of the dated intrusions is 1.59 to 1.54 Ga. Temporally part of this group, but emplaced east of the Wiborg massif, are the Salmi and Sotjärvi plutons.

All the 1.65 to 1.54 Ga AMCG intrusions were emplaced into Svecofennian crust with the exception of the Salmi and Sotjärvi bodies, which are located in Proterozoic-reworked Archean crust.

1.51 to 1.43 Ga

Between 1.51 and 1.43 Ga, the focus of activity shifted farther west, and included widely scattered felsic, and locally mafic, magmatism. Included here are granite plutons in southeast Sweden (Åberg, 1988), gabbroic intrusions in southwest Sweden (Åhäll *et al.*, 1990), alkaline intrusions in the Bamble Sector (Starmer, 1990), felsic volcanic rocks in the Telemark Sector (Dahlgren and Heaman, 1991) and granitic and gabbroic intrusions in the Western Gneiss Region (Tucker *et al.*, 1990). A period of high-grade metamorphism at this time has been identified in southwesternmost Norway (Corfu, 1980), and it is possible that many of the intrusions farther east are inboard magmatic effects linked with that event. Mafic intrusive activity at 1.43 Ga, which was apparently widespread in both Laurentia and Baltica, terminates these events.

1.43 to 1.27 Ga

Following the 1.43 Ga mafic magmatism referred to above, sporadic and scattered granite emplacement continued across southern Scandinavia, especially in southeast and southwest Sweden, southeast Norway and in the Bamble Sector in southern Norway. This stage of crustal evolution in Baltica and Laurentia also ended with widespread mafic intrusive activity, which occurred at 1.27 Ga (cf. Gower *et al.*, 1990).

METALLOGENY

Mineralization related to the anorogenic suites can be divided into in two groups:

(i) Sn–Be–W–Pb–Zn mineralization associated with specialized granites, and

(ii) Ti–V mineralization associated with fractionated mafic intrusions.

Sn–Be–W–Pb–Zn Mineralization

This metallogenic type is associated with the 1.65 to 1.54 Ga rapakivi massifs. Mineralization occurs in the Wiborg, Vehmaa, Åland and Salmi massifs and in the Eurajoki, Suomenniemi and Ahvenisto satellite intrusions. The only economic deposits are at Pitkäranta in the Salmi massif, where mining started in 1842 (Haapala, 1977).

The mineralization is associated with the youngest equigranular or porphyritic rapakivi granite types. It is concentrated in pegmatitic veins and pockets, and in greisen and quartz veins. The ore minerals include cassiterite, beryllium minerals (beryl, genthelvite, bertandite), wolframite, sphalerite, chalcopyrite, galena, molybdenite, arsenopyrite and gahnite. These are associated with a wide range of accessory phases, such as topaz, tourmaline, fluorite, monazite and bastnaesite, in addition to more common silicate phases such as quartz, mica and chlorite (Haapala, 1988). The Pitkäranta deposits are somewhat different, in that they occur in limestone–dolomite-skarn layers in the exocontact zone of the Salmi massif, parts of which are probably situated below and very close to the deposits, although the main massif is up to 4 km away on the surface. Despite the mineralization not being in the massif, there seems little reason to doubt a genetic link between the mineralization and the rapakivi intrusion (Haapala, 1977) and, like similar mineralization elsewhere, to interpret the ore minerals as having been precipitated from late-stage residual fluids. An important

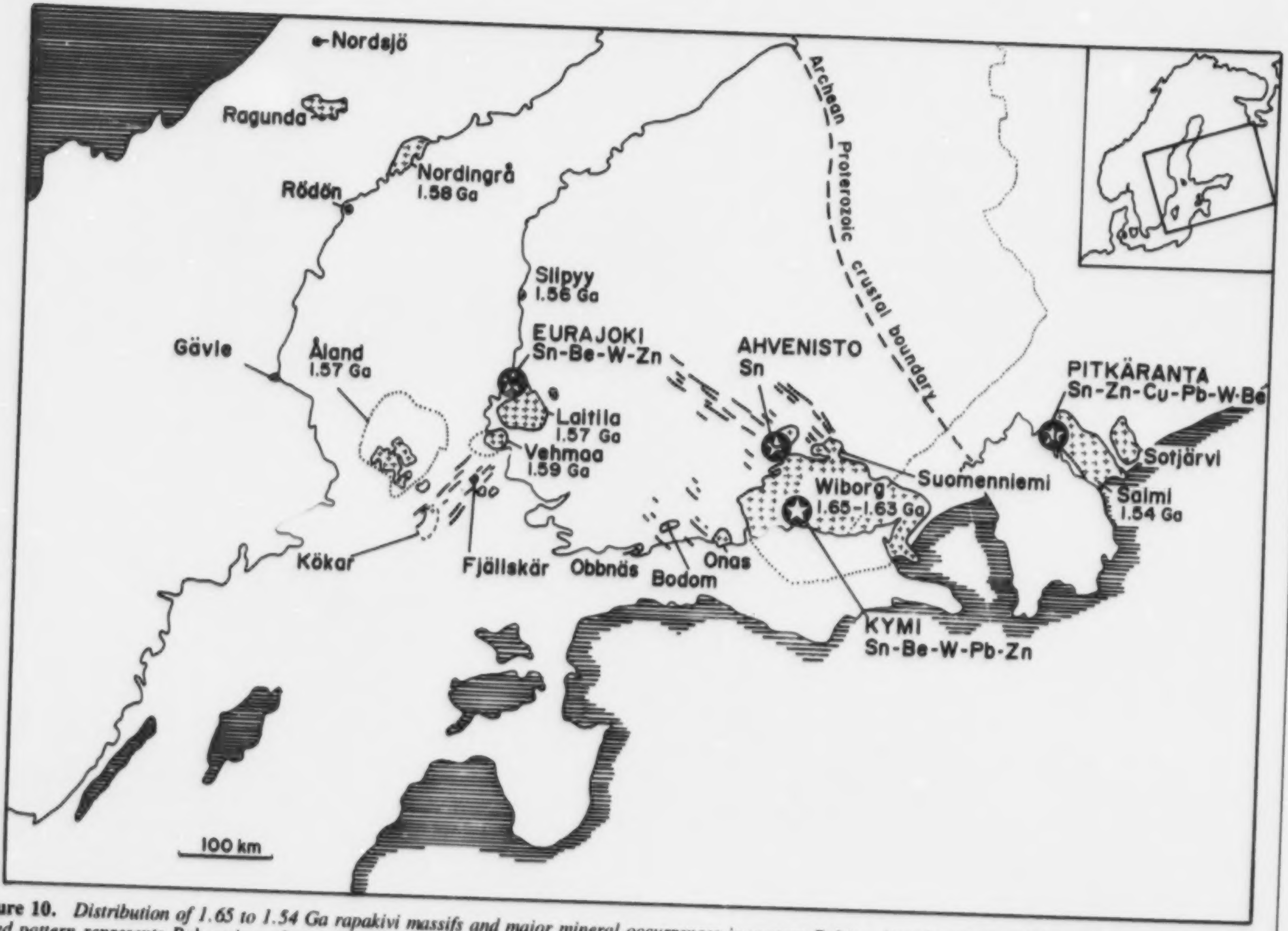
Nordsjö
Ragunda
Nordingrå
1.58 Ga
Rödön
Silpyy
1.56 Ga
Gävle
Åland
1.57 Ga
EURAJOKI
Sn-Be-W-Zn
Laitila
1.57 Ga
Vehmaa
1.59 Ga
Kökar
Fjällskär
Obbnäs
Bodom
Onas
AHVENISTO
Sn
Archean Proterozoic crustal boundary
Wiborg
1.65-1.63 Ga
Suomenniemi
KYMI
Sn-Be-W-Pb-Zn
PITKÄRANTA
Sn-Zn-Cu-Pb-W-Be
Sotjärvi
Salmi
1.54 Ga
100 km

Figure 10. *Distribution of 1.65 to 1.54 Ga rapakivi massifs and major mineral occurrences in eastern Baltica. Based on Rämö (1991) and Lundqvist (1990). Lined pattern represents Paleozoic rocks.*

difference between the Pitkäranta deposits and those of other rapakivi massifs is that the metals source for the Pitkäranta deposits was Archean rather than Svecofennian (Sundblad, 1991b, Rämö, 1991).

Ti–V Mineralization

The Ti–V deposits occur in various places in Baltica associated with mafic intrusions emplaced during the anorogenic period. The only two of major importance are those at Smålands Taberg in southern Sweden and at Ulvön in central east Sweden (Figure 11). The mineralization in the Smålands Taberg deposit consists of rich concentrations in an olivine-bearing metagabbro emplaced between 1550 and 1500 Ma. It comprises titanomagnetite, with which minor pyrite, pyrrhotite, pentlandite and chalcopyrite are associated. Although large (150 million tonnes), the grade of 6 percent TiO_2 and 0.3 percent V_2O_5, is too low for the deposit to be economic (Grip, 1978). The Ulvön Ti–V deposit is much younger, being part of the widespread ca. 1270 Ma magmatic event that affected both Baltica and Laurentia. The ore minerals are titanomagnetite with ilmenite and ulvöspinel and they occur at several levels in an olivine dolerite intrusion (Larson and Magnusson, 1976). As the layers are rarely thicker than 2 m (6 m thick including disseminated zones) and the deposit is relatively small (20 million tonnes), it is not currently economic. The grade is slightly higher than that at Smålands Taberg, achieving 8 percent TiO_2 and 0.35 percent V_2O_5 (Grip, 1978).

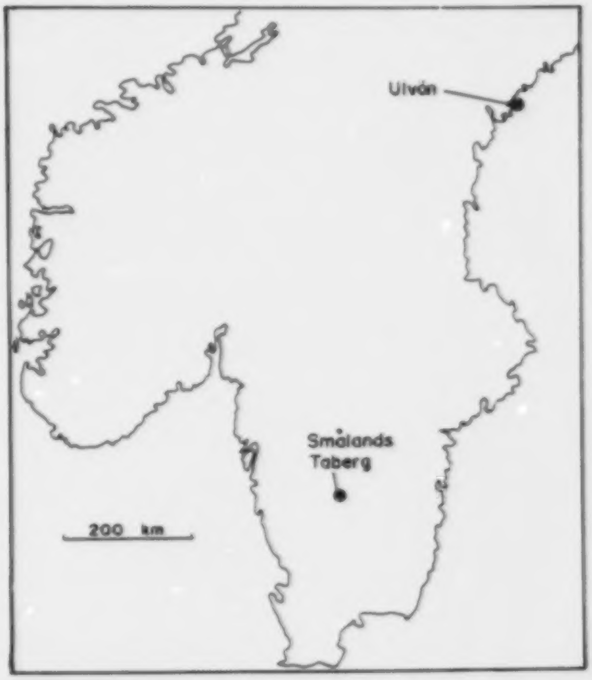

Figure 11. *Locations of the Smålands Taberg and Ulvön Ti–V deposits (Grip, 1978).*

LAURENTIAN ANALOGUES (1.65 to 1.27 Ga)

Labrador is well known for its mid-Proterozoic anorogenic intrusive and associated volcanic rocks. These include the 1.64 Ga Mealy Mountains Intrusive Suite, the ca. 1.48 Ga Pinware terrane granites, the 1.45 Ga Harp Lake, Michicamau, and Mistastin intrusions, the 1.43 Ga Michael–Shabogamo gabbros, the 1.33 to 1.27 Ga Nain Plutonic Suite, the 1.33 Ga Red Wine Intrusive Suite and genetically related Letitia Lake Group and the 1.27 Ga Flowers River Suite (Figure 12).

The spectrum of ages for the Labrador intrusions closely matches that of Baltica and it would be possible to make direct chronological comparisons with their Baltica analogues. A much more important metallogenic factor than precise time of emplacement within the anorogenic period, however, is the exposed structural level. The Sn–Be–W–Pb–Zn mineralization occurs in the uppermost levels of granitoid intrusions emplaced at high crustal levels, whereas Ti–V mineralization is found near the base of mafic intrusions.

Some of the highest level granitoid intrusions must have been near surface because some of the magma escaped to produce volcanic rocks still present in the region. Thus the Middle Proterozoic (at ca. 1.30 Ga) and present-day surfaces are essentially identical in some parts of Labrador. Examples in Labrador include the Red Wine Intrusive Suite (together with the spatially and genetically associated Letitia Lake Group), the Flowers River Suite, and the Strange Lake intrusion (cf. Hill and Miller, 1990). All of these units host mineralization resembling that found in the Finnish rapakivi granites (cf. Miller, 1988, 1990). Larger felsic bodies that ascended to high crustal levels include the Notakwanon, Makhavinekh and Umiakovik intrusions, which all form part of the Nain Plutonic Suite. These are not known to be mineralized, but their close similarity to ore-mineral-bearing rapakivi granites in Baltica has been emphasized by Ryan (1991). Compositionally similar, but somewhat older rocks include the 1.44 Ga Mistastin intrusion, fringes of the 1.45 Ga Harp Lake and Michicamau bodies and part of the Mealy Mountains Intrusive Suite. The Mealy Mountains intrusion is easily overlooked as having any potential for granophile mineralization, particularly as it occurs in the Grenville Province, and the best-known northern part is dominantly anorthositic and probably represents an upthrust, deeper level of the intrusion. It seems likely that higher intrusive levels will be identified south of the areas mapped at present. Despite the region being in the Grenville Province, there is good evidence that the Mealy Mountains Intrusive Suite was never deeply buried during Grenvillian orogenesis (Emslie *et al.*, 1984).

Two further points are worth making. First, although the largest bodies provide the biggest tracts of prospective terrain, they may not represent the best targets. Given that mineralization is concentrated near the roof of intrusions, better prospects would be granites that did not quite reach the earth's surface, and did not vent to it. If exposed, such granites would appear quite small and seemingly insignificant (e.g., Strange Lake). Second, in Baltica, the best

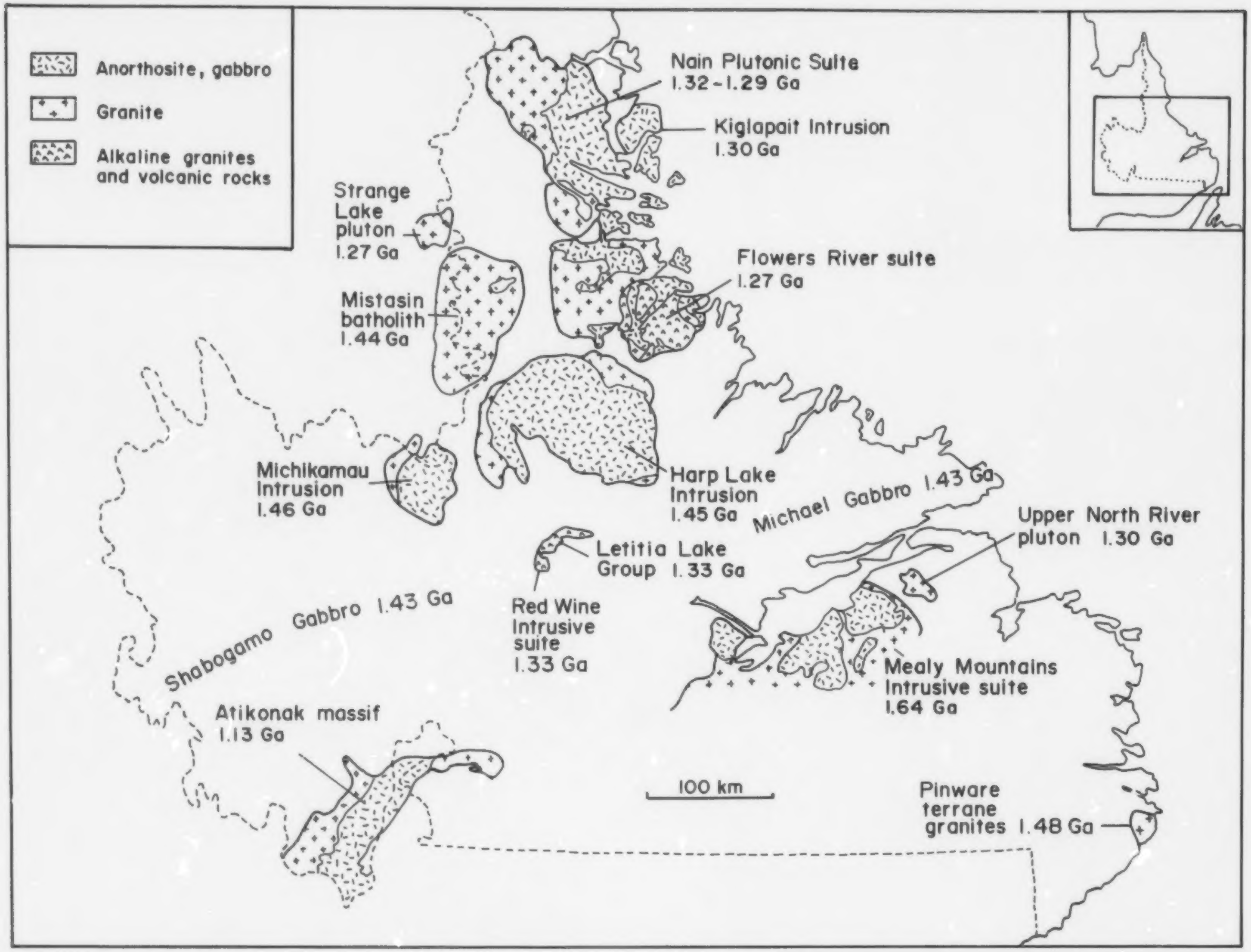
Anorthosite, gabbro
Granite
Alkaline granites and volcanic rocks
Nain Plutonic Suite 1.32-1.29 Ga
Kiglapait Intrusion 1.30 Ga
Strange Lake pluton 1.27 Ga
Flowers River suite 1.27 Ga
Mistasin batholith 1.44 Ga
Michikamau Intrusion 1.46 Ga
Harp Lake Intrusion 1.45 Ga
Michael Gabbro 1.43 Ga
Upper North River pluton 1.30 Ga
Letitia Lake Group 1.33 Ga
Red Wine Intrusive suite 1.33 Ga
Shabogamo Gabbro 1.43 Ga
Mealy Mountains Intrusive suite 1.64 Ga
Atikonak massif 1.13 Ga
100 km
Pinware terrane granites 1.48 Ga

Figure 12. *Distribution of major 1.65 to 1.27 Ga anorogenic intrusions in Labrador. Location of individual intrusions of Shabogamo and Michael gabbros are not shown. The 1.13 Ga Atikonak massif (mentioned later in the text) is also shown.*

mineralization is associated with an intrusion (Salmi) that was emplaced into crust having a long and active earlier history; by analogy, the best prospects in Laurentia should be associated with anorogenic intrusions emplaced into Archean crust reworked during (repeated) Proterozoic orogenesis.

Where there is potential for high-level Sn–Be–W–Pb–Zn mineralization, then, at first thought, it would seem fortuitous to expect Ti–V mineralization in rocks of similar age, remembering that such mineralization is most probable in the lower level of mafic intrusions. If, however, the mafic magmas reached moderately high levels prior to fractionation, then both types of mineralization might conceivably be found in the same area. Also, it can only be established that the earth's surface at ca. 1.30 Ga (and north of the Grenville front) was the same as the present earth's surface; pre-1.30-Ga intrusions were uplifted and deeper levels exposed (Hill, 1991). Such a situation, admittedly, diminishes prospects for Sn–Be–W–Pb–Zn mineralization in the older intrusions, but it might improve their chances for hosting near-surface Ti–V ore bodies. In any case, Ti–V mineralization is known in several of the above listed mafic intrusions, although, like their Baltica counterparts, the occurrences are too small and(or) low grade to be exploited.

PRE-SVECONORWEGIAN/ SVECONORWEGIAN–PRE-GRENVILLIAN/ GRENVILLIAN

> *The recognition of large amounts of auriferous quartz veins in southwestern Scandinavia has been one of the pleasant suprises of this work.* (Gaál and Sundblad, 1990).

GEOLOGICAL SETTING

The period between 1.27 and 0.95 Ga in both Baltica and Laurentia can be subdivided into two orogenic events separated by an anorogenic stage that lasted from ca. 1.20 to 1.10 Ga (cf. Gower *et al.*, 1990). In southwest Baltica, between 1.27 and 1.20 Ga, there was widespread granitoid plutonism accompanied by high-grade metamorphism. This event is best documented in southwest Sweden, where several crustally derived granite intrusions were emplaced at about 1.25 Ga (Lindh and Holme, 1989). In southeast Sweden, syenite to granite intrusions were emplaced between 1.23 and 1.20 Ga (Johansson, 1990; Hansen and Lindh, 1991). In southern Norway, although the timing of similar granitoid plutonism and metamorphism is not so precisely constrained, the pattern of events appears to be somewhat similar and the events are grouped together as early Sveconorwegian (Falkum, 1985; Smalley *et al.*, 1988; Starmer, 1990, 1991).

Following early Sveconorwegian orogenesis, there was widespread mafic extrusive and intrusive activity and some sedimentation, probably in a rift-related setting (cf. Atkin and Brewer, 1990). The focus of this activity in Norway was the Telemark Sector where clastic sediments of the Seljord Group were deposited, and were overlain by bimodal volcanic rocks (1.16 Ga; Dahlgren and Heaman, 1991) and associated sediments of the Bandak Group. At the same time, the Seljord Group was intruded by mafic sills. Similar mafic intrusive activity also occurred in the adjacent Bamble Sector. In southwest Sweden, the (probably rift-controlled) Dalsland Group was deposited, and the 1.18 Ga Protogine Zone dykes were emplaced (cf. Johansson, 1990). Some granitic intrusive activity also took place, peaking at ca. 1.14 Ga in both southern Norway and southwest Sweden (cf. Persson *et al.*, 1987).

The renewed orogenesis that started at ca. 1.10 Ga resulted in widespread deformation and high-grade metamorphism and was accompanied by the emplacement of a wide range of syn- to post-kinematic high-K calc-alkaline to alkali-calcic granitoid rocks (Eliasson and Schöberg, 1991), mafic intrusions, and, at the end of the period in southwest Norway, the Rogaland AMCG suite (Duchesne and Michot, 1987).

The anorogenic stage can be linked with rifting of Baltica away from easternmost Laurentia (probably east Greenland). The final orogenesis is interpreted here to correlate with collision between Baltica and southeastern Laurentia (southern Labrador and eastern Quebec), after its rotation through about 90° (Gower, 1990). The rotation is dated between 1.10 and 1.00 Ga from paleomagnetic evidence (Stearn and Piper, 1984). After ca. 1.05 Ga, therefore, southwest Baltica would have been juxtaposed against southeastern Laurentia, and they should have directly complementary histories.

METALLOGENY

The following metallogenic environments are represented in the Sveconorwegian Province:

(i) Cu–Mo–Au mineralization hosted by rift-related supracrustal rocks of the Telemark Supergroup,

(ii) Mo mineralization linked with syn- to late-tectonic granitoid rocks,

(iii) Au–Cu–(Pb–Zn–Ag) mineralization associated with late- to posttectonic veins, and

(iv) Ti–Fe–Ni mineralization in the Rogaland anorthosite massif.

Recent information concerning the first two of these metallogenic environments is rather sparse, and most of the information presented here for them is summarized from Bugge (1978).

Cu–Mo–Au Mineralization

This type of mineralization is found in an area measuring roughly 150 by 50 km in the Telemark Sector in southern Norway (Figure 13). Probably similar, but commercially

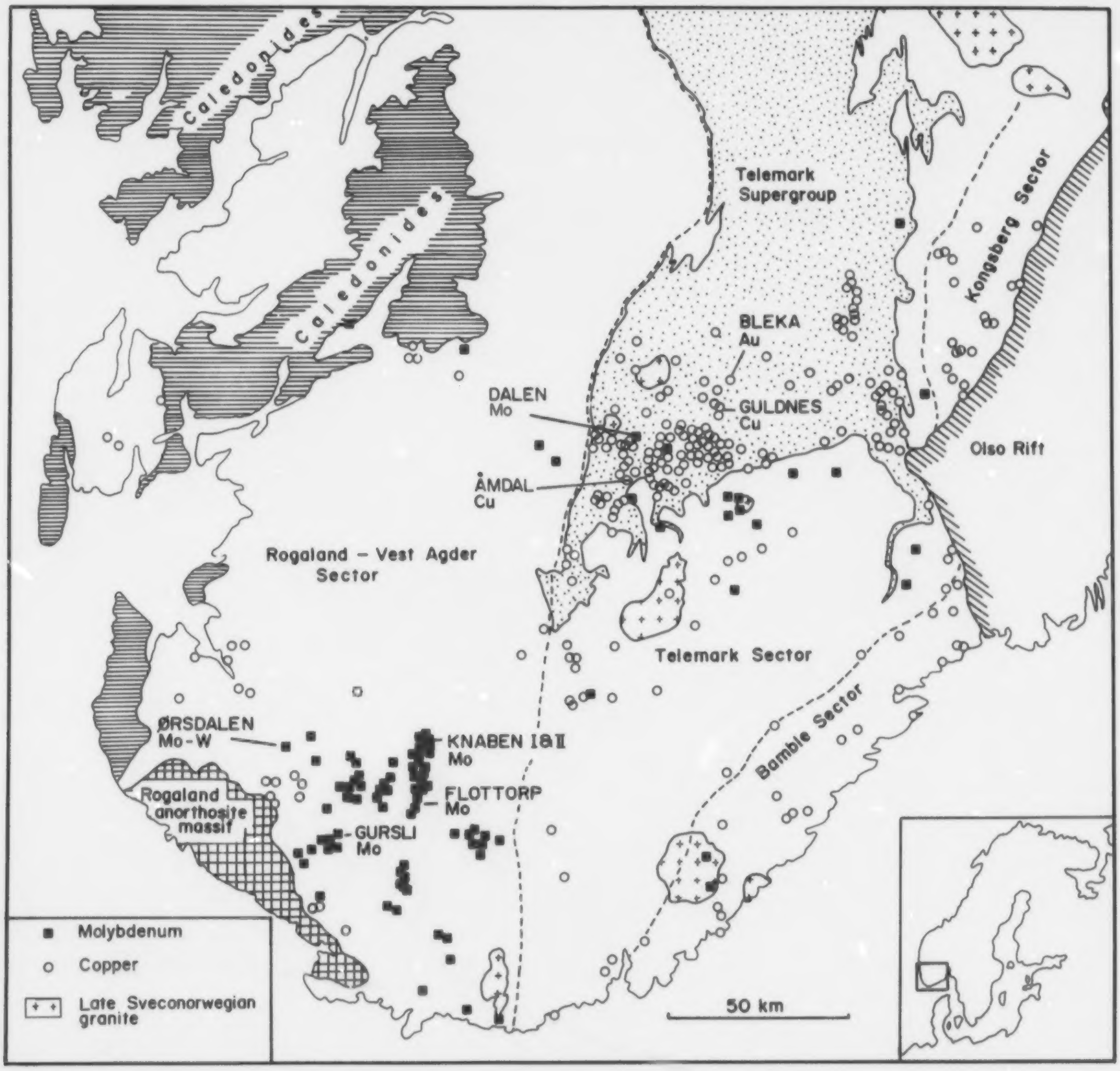
Caledonides
Caledonides
Telemark Supergroup
BLEKA Au
GULDNES Cu
DALEN Mo
ÅMDAL Cu
Kongsberg Sector
Oslo Rift
Rogaland – Vest Agder Sector
Telemark Sector
Bamble Sector
ØRSDALEN Mo-W
KNABEN I&II Mo
FLOTTORP Mo
GURSLI Mo
Rogaland anorthosite massif
■ Molybdenum
○ Copper
+ + Late Sveconorwegian granite
50 km

Figure 13. *Distribution of Cu and Mo mineralization in the Sveconorwegian Province in southwest Norway. Based on data compiled by Torske (1976) and Sigmond (1985).*

insignificant mineralization is known within the Dalsland Group in southwest Sweden. The deposits in the Telemark Sector supported several mining operations from the beginning of the sixteenth century until 1945. The principal commodity exploited was Cu, but mining was also carried out for Mo, As, Au and Bi.

The deposits are hosted by rocks of the Telemark Supergroup, especially the Seljord and Bandak groups. The deposits occur in irregular pods and lenses in quartz veins, that may be up to a few metres wide and several hundred metres long. A wide variety of minerals have been found, but the most important, economically, are chalcopyrite, bornite, digenite, chalcocite, molybdenite, galena, sphalerite, arsenopyrite and electrum. Silver and bismuth minerals are found in most of the Cu deposits. The ore minerals are associated with quartz, calcite, micas, feldspars, chlorite, fluorite and tourmaline. Pyrite is only found as an accessory mineral in chalcopyrite-bearing deposits and pyrrhotite is rare.

The mineralization in the Dalsland Group (Figure 14) consists of stratabound Cu impregnations in a calcareous phyllite sandwiched between two quartzite beds (Grip, 1978; Johansson, 1985).

The Telemark deposits were originally interpreted as hydrothermal fissure fillings genetically related to granite emplacement or to mobilization during high-grade

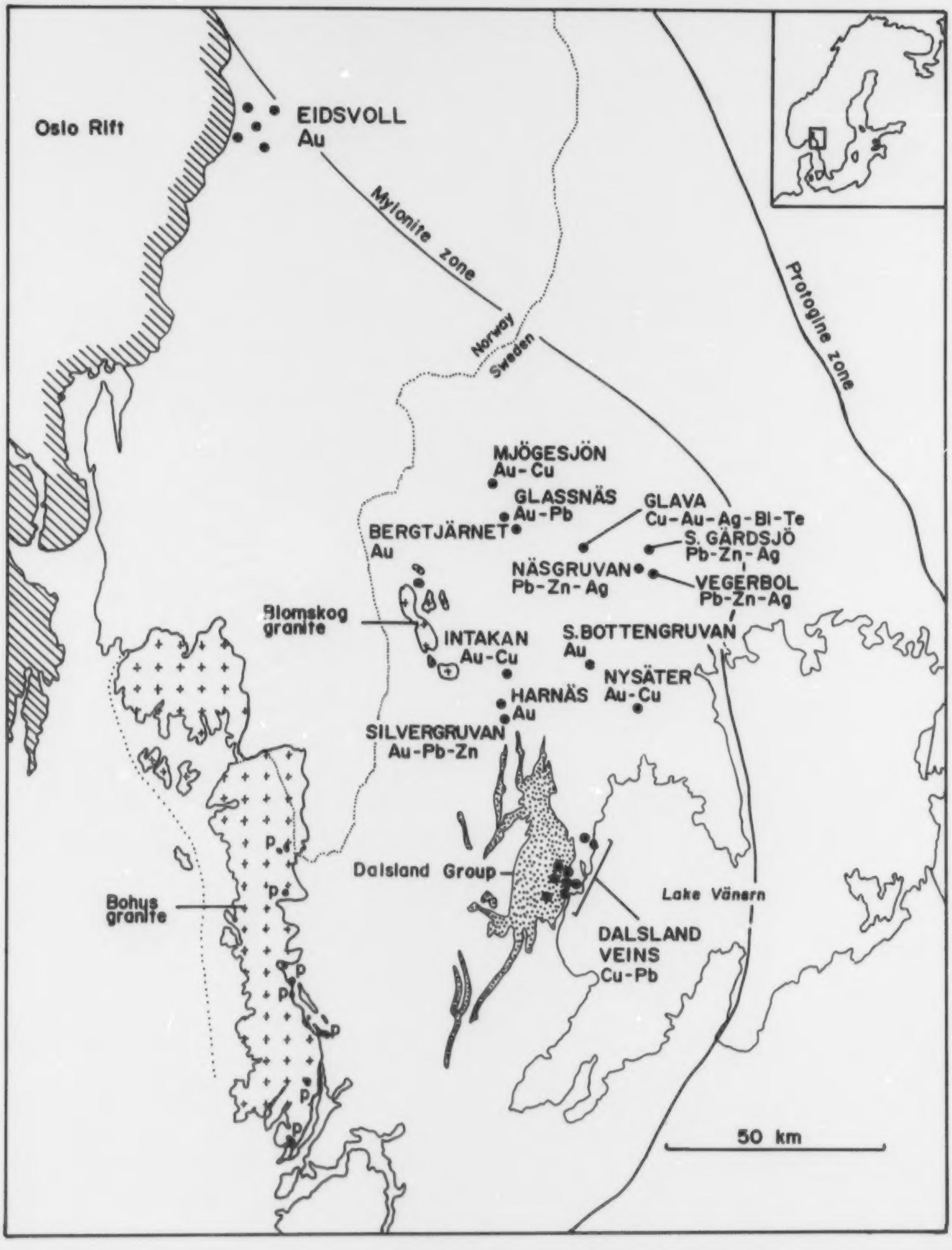
Oslo Rift
EIDSVOLL
Au
Mylonite zone
Norway
Sweden
Protogine zone
MJÖGESJÖN
Au-Cu
GLASSNÄS
Au-Pb
GLAVA
Cu-Au-Ag-Bi-Te
BERGTJÄRNET
Au
S. GÄRDSJÖ
Pb-Zn-Ag
NÄSGRUVAN
Pb-Zn-Ag
VEGERBOL
Pb-Zn-Ag
Blomskog granite
INTAKAN
Au-Cu
S.BOTTENGRUVAN
Au
NYSÄTER
Au-Cu
HARNÄS
Au
SILVERGRUVAN
Au-Pb-Zn
Dalsland Group
Lake Vänern
Bohus granite
DALSLAND
VEINS
Cu-Pb
p
50 km

Figure 14. *Distribution of vein-type mineralization in southeast Norway and southwest Sweden. Compiled from Alm (1991) and Johansson (1985), p—pegmatite at the eastern margin of the Bohus granite.*

metamorphism. The spatial association with the supracrustal rocks cannot be denied, however, and it has been more recently suggested that the mineralization was originally syngenetic, and that the epigenetic imprint was subsequently stamped onto the deposits during metamorphism and granitoid emplacement.

Mo Mineralization

In contrast to the Cu-dominant mineralization of the Telemark Sector, Mo mineralization is mainly found farther west, in the Rogaland–Vest Agder Sectors (Sigmond, 1985). The Knaben deposits are the most important in this region and were mined over a period of nearly 100 years. The largest was the Knaben II deposit from which 8.6 million tonnes of ore was extracted, averaging 0.21 percent MoS_2. The mine is no longer in production, but proved reserves of 2.5 million tonnes of ore, averaging 0.15 percent MoS_2 remain. Smaller exploited deposits include the Ørsdalen Mo-W mine, and the Flottorp and Gursli mines (Figure 13).

Mo mineralization occurs in amphibolitic gneiss, albitized granite, aplite, pegmatite, and quartz veins and lenses. Most of the dykes and veins are small, but production from the Knaben II deposit exploited a quartz vein 30 to 80 m wide and 1500 m long. In the minor granitoid intrusions and quartz veins, the grade is commonly high, but drops off rapidly in disseminated zones in granite and amphibolitic gneiss. In addition to molybdenite, pyrite, chalcopyrite, pyrrhotite, scheelite, sphalerite and bismuth minerals may be present. Associated minerals include quartz, muscovite, calcite and fluorite.

The deposits have been compared to Cu–Mo porphyry deposits elsewhere (Torske, 1976; Bugge, 1978), although there are differences, particularly in the Norwegian examples being smaller and lacking associated volcanic rocks. The granitoid intrusions with which the late-stage dykes and veins are linked, are interpreted to be of deep crustal origin.

Au–Cu–(Pb–Zn–Ag) Mineralization

In contrast to the dearth of up-to-date information available on polymetallic mineralization in the Norwegian part of the Sveconorwegian Province, there has been sustained recent interest in vein-type mineralization (Figure 14) in southwest Sweden (Johansson, 1985; Lundegårdh, 1989; Gaál and Sundblad, 1990; Alm, 1991). Some of the larger veins have been mined intermittently over the last 200 years, but none is being exploited at present. The most important past producers were at Eidsvoll in southeast Norway (for Au), and at Glava in southwest Sweden (for Cu).

The veins occur in a broad belt west of the mylonite zone (Ihlen, 1986), a region representing a shallower level of Sveconorwegian orogenesis than the rocks immediately east of the mylonite zone. In addition to the veins located in Figure 14, Au has been discovered in Halland, south of Gothenburg (Arbetet newspaper, 16 May 1991). Although in a rather different tectonic setting, the Bleka Au mine in the Telemark Sector (Figure 13) may also belong to this metallogenic group.

Most of the mineralization is hosted by felsic to intermediate gneisses, mafic rocks and porphyries of Gothian–Kongsbergian age (see earlier), but some of the southerly veins are located within rocks of the Dalsland Group, which constrains their age to less than ca. 1200 Ma. A better indication of age is provided by Johansson (1985), who concluded from Pb isotopic data that the best estimate for the time of mineralization was 1010 ± 90 Ma for the veins from both Dalsland and Värmland.

The veins are typically small (rarely wider than 1 to 2 m and rarely longer than 100 m), steeply dipping, and are probably fault controlled. A wide range of minerals is present, but the most common are pyrite, chalcopyrite, bornite, covellite, galena and sphalerite (Johansson, 1985; Alm, 1991). Alm also concluded that the Cu-rich, Au-bearing veins preferentially carried Te-, Se- and Bi-bearing phases.

Johansson (1985) suggested that the veins could have formed during either Dalslandian (1000 Ma) metamorphism or associated with the emplacement of the Bohus (0.92 Ga) and Blomskog granites (located on Figure 14). The metals were introduced along with hydrothermal fluids, and isotopic data indicate that they probably originated in rocks formed during Gothian–Kongsbergian orogenesis.

Ti–Fe–Ni Mineralization

The Ti–Fe–Ni mineralization in the Sveconorwegian Province can be divided into two groups, (i) currently uneconomic deposits associated with various host rocks and scattered throughout the Sveconorwegian Province, but mainly concentrated in the Bamble Sector, and (ii) large deposits associated with the Rogaland anorthosite massif in southwest Norway.

Deposits of the first group include ilmenite deposits in layered gabbros and amphibolite, skarn-type ores associated with calcareous metasedimentary gneisses, apatite-bearing ores associated with quartzofeldspathic gneisses and fissure vein deposits in olivine gabbros consisting of magnetite with quartz, rutile and hematite. The host rocks have ages between ca. 1.70 and 1.10 Ga, and the deposits, although mined in the past, are generally too small and low grade to attract current interest.

In contrast, the deposits associated with the Rogaland anorthosite massif (Figure 15) are among the largest and most important of their kind in Europe. Production started in the late 19th century from the Blåfjell and Koldal-Anker mines, and in 1918 from the Storgangen deposit. Most important at the moment, however, is the huge Tellnes deposit, which, although in the same district, was not discovered until 1954. This deposit, from which production commenced in 1960, has reserves of over 300 million tonnes averaging 18 percent TiO_2 (Bugge, 1978), and currently accounts for more that 13

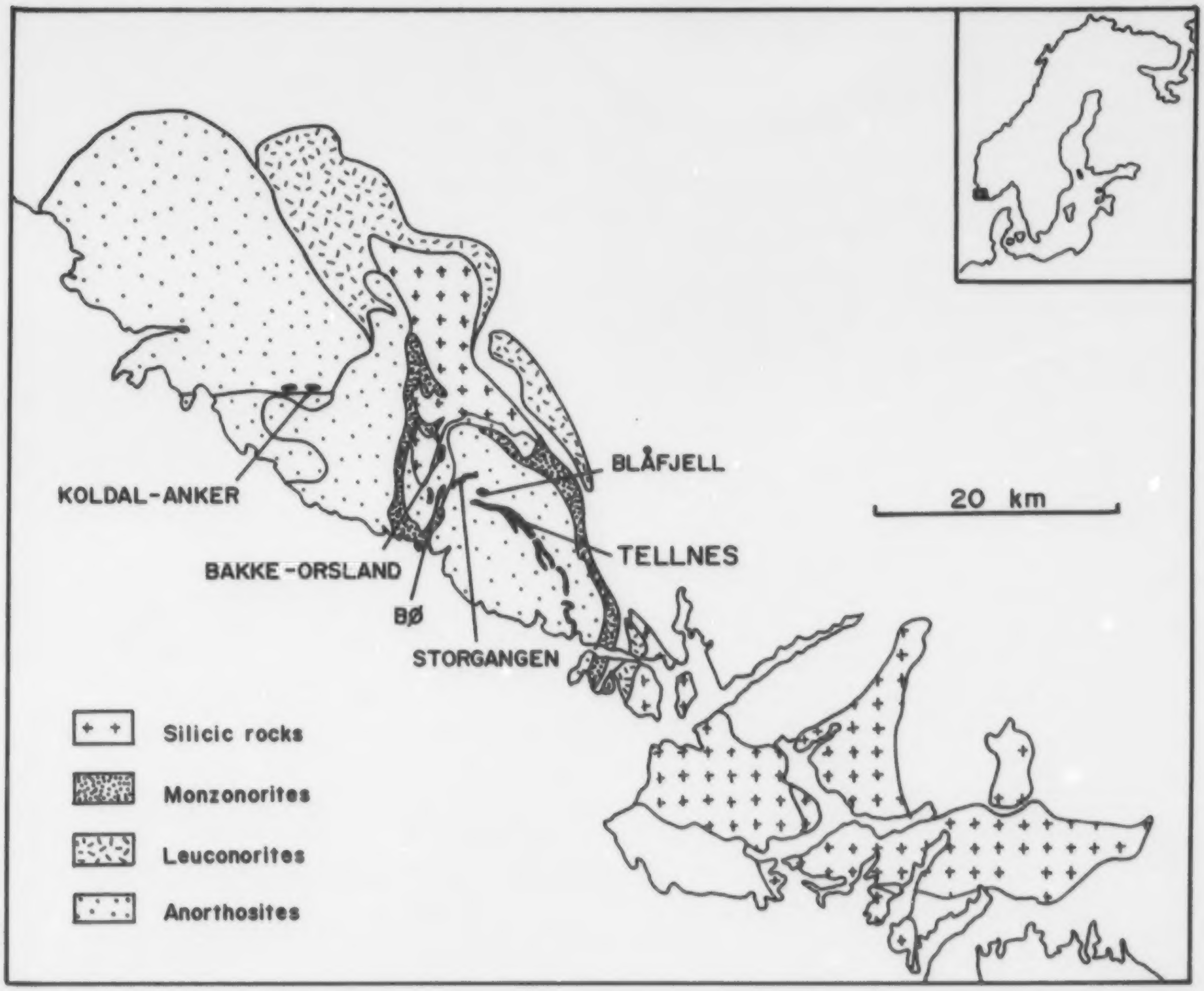

Figure 15. *Major rock units and location of Ti–Fe–Ni deposits in the Rogaland anorthosite massif (after Bugge, 1978).*

percent of the world supply of TiO_2 pigment (Wilmart *et al.*, 1989).

The Tellnes orebody is an ilmenite norite lens intruded into the surrounding anorthosite. The deposit is 400 m thick and 2.5 km long and extends at both ends as a 5- to 10-m-thick dyke that can be traced for 4 km and 10 km in northwest and southeast directions, respectively. In addition to ilmenite, the ore contains plagioclase, bronzite and olivine, together with lesser clinopyroxene, apatite, biotite, magnetite, and complex Fe–Ni–(Co)–Cu sulphides. The ilmenite norite and more evolved felsic units in the same region are interpreted to be the product of fractional crystallization (Wilmart *et al.*, 1989).

LAURENTIAN ANALOGUES (1.27 to 0.95 Ga)

The most obvious analogue to the Telemark Supergroup would seem to be the Seal Lake Group as both sequences formed in similar tectonic settings and show similar mineralization styles. The two sequences cannot be strictly compared, however, because the Seal Lake Group was probably deposited at least 70 million years earlier (and is better equated with Jotnian rocks). Nevertheless, in view of the fact that, in both regions, the mineralization has been interpreted to be epigenetic (cf. Wilton, 1989 for the Seal Lake Group) and related to Grenvillian–Sveconorwegian orogenesis, comparison may be justified.

An alternative Laurentian analogue to the Telemark Supergroup is the Wakeham Supergroup in eastern Quebec (Figure 16). The lower Aguanus River Group has an age of 1.27 Ga, but the overlying Davy Lake Group has not been dated (Bourne, 1991). The Davy Lake Group is a good analogue for the Telemark Supergroup, as the Wakeham Supergroup is situated well within the Grenville Province (as is the Telemark Supergroup in the Sveconorwegian Province) and the Davy Lake Group (which must be older than 1.11 Ga; Bourne, 1991), could easily have the same 1.16 Ga age

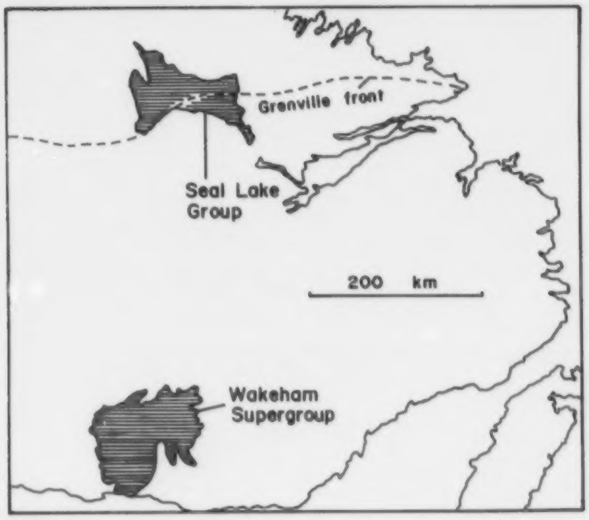

Figure 16. *Locations of Seal Lake Group and Wakeham Group in eastern Laurentia.*

as the Bandak Group. Of particular interest to mineral exploration in Labrador is the tectonic relationship between the Seal Lake Group and the lower part of the Wakeham Supergroup, and whether similar rocks (or mineralized linking structures) remain undiscovered in the poorly mapped intervening area in southernmost Labrador.

There are no known examples in Labrador comparable to either the Mo mineralization in the western Sveconorwegian Province or the Au–Cu–(Pb–Zn–Ag) mineralization in the eastern Sveconorwegian Province. Rather than this demonstrating a weakness in the Baltica–Laurentia correlative approach, given all the other geological correlations that exist, lack of documented mineralization offers tantalizing mineral-exploration possibilities for southern Labrador. It is not difficult to understand why mineralization might have escaped detection–large areas of southernmost Labrador are poorly exposed and unmapped, even at 1:250 000 reconnaissance scale.

Mineralization in the Sveconorwegian Province is interpreted to be an integral part of Sveconorwegian orogenesis and associated late- to posttectonic granite emplacement. It is only in recent years that it has been recognized that similar granites (Figure 17) are widespread in southern Labrador (Gower *et al.*, 1991). They are not presently known to contain Mo or Au mineralization. The granites locally contain U mineralization (Hauseux, 1977), they are generally enriched in large ion lithophile (LIL) elements, and they have radio-element and lake-sediment granophile anomalies (e.g., Gower *et al.*, 1985). The link between radio-element enrichment and other mineralization is important as, in the Sveconorwegian Province, radio-element-anomalous granitoid rocks clearly correlate with the distribution of Mo mineralization in the Rogaland–Vest Agder Sector (Sigmond, 1985) and vein mineralization in southwestern Sweden has been temporally linked to the anomalously radioactive Bohus and Blomskog granites (Gaál and Sundblad, 1990).

There are no direct temporal Laurentian analogues for the Rogaland anorthosite massif in Labrador, and it seems unlikely that any will be found. As noted earlier, however, temporal correlation is probably less important than tectonic setting in the case of Ti–Fe–V mineralization in anorthosites. The commercial Lac Tio deposits in the Allard Lake anorthositc and the unexploited Magpie Mountain deposit in eastern Quebec are examples of similar deposits in slightly older (1.13 Ga) anorthosite massifs in the eastern Grenville Province (Sangster *et al.*, *in press*). The Atikonak massif, in southwest Labrador, is a temporal equivalent of the Lac Allard massif.

A final word of caution is necessary with respect to Sveconorwegian–Grenvillian metallogenic correlations. Although various metallogenic provinces seem to be fairly clearly outlined in the Sveconorwegian Province, the identification of similar regions in southern Labrador is not straightforward, given that the orientation of Baltica changed considerably during this period. Thus the pattern, from east to west, of 'Sveconorwegian front–Au metallogenic zone–Mo metallogenic zone', need not be matched by a similar north-to-south progression from the Grenville front. Such a comparison might be justified if these metallogenic characteristics were acquired prior to rotation, and simply modified during subsequent collision, but if the metallogenic signatures postdate ca. 1.00 Ga, then they will be a product of the collision and the above correlation would be invalid.

FINAL COMMENTS

This article has examined mineral occurrences in Baltica formed during a 1-billion-year period when Laurentia and Baltica experienced a common geological history. Geological comparisons have been made largely on a temporal basis, although it is recognized that the identification of appropriate tectonic environments, irrespective of their age, is the most widely established predictive metallogenic strategy. Nevertheless, by adopting a temporal correlation approach (incorporating tectonic comparisons), several new exploration possibilities have emerged. Some of these are in line with established thinking, but some have not been previously considered and, hopefully, will develop into new and successful exploration models for the region.

This article has also emphasized, for rocks of several different ages, the importance of multistage reconcentration of metals in the crust. Such a process is heavily dependent on either superimposed orogenesis or anorogenic magmatism within pre-existing crust. In this regard, it is vital to know when the crust separated from the mantle, and to understand its subsequent geological history. Such information can be obtained by combining Sm–Nd and U–Pb isotopic techniques, and these investigations should routinely complement every regional mapping or regional mineral-exploration project.

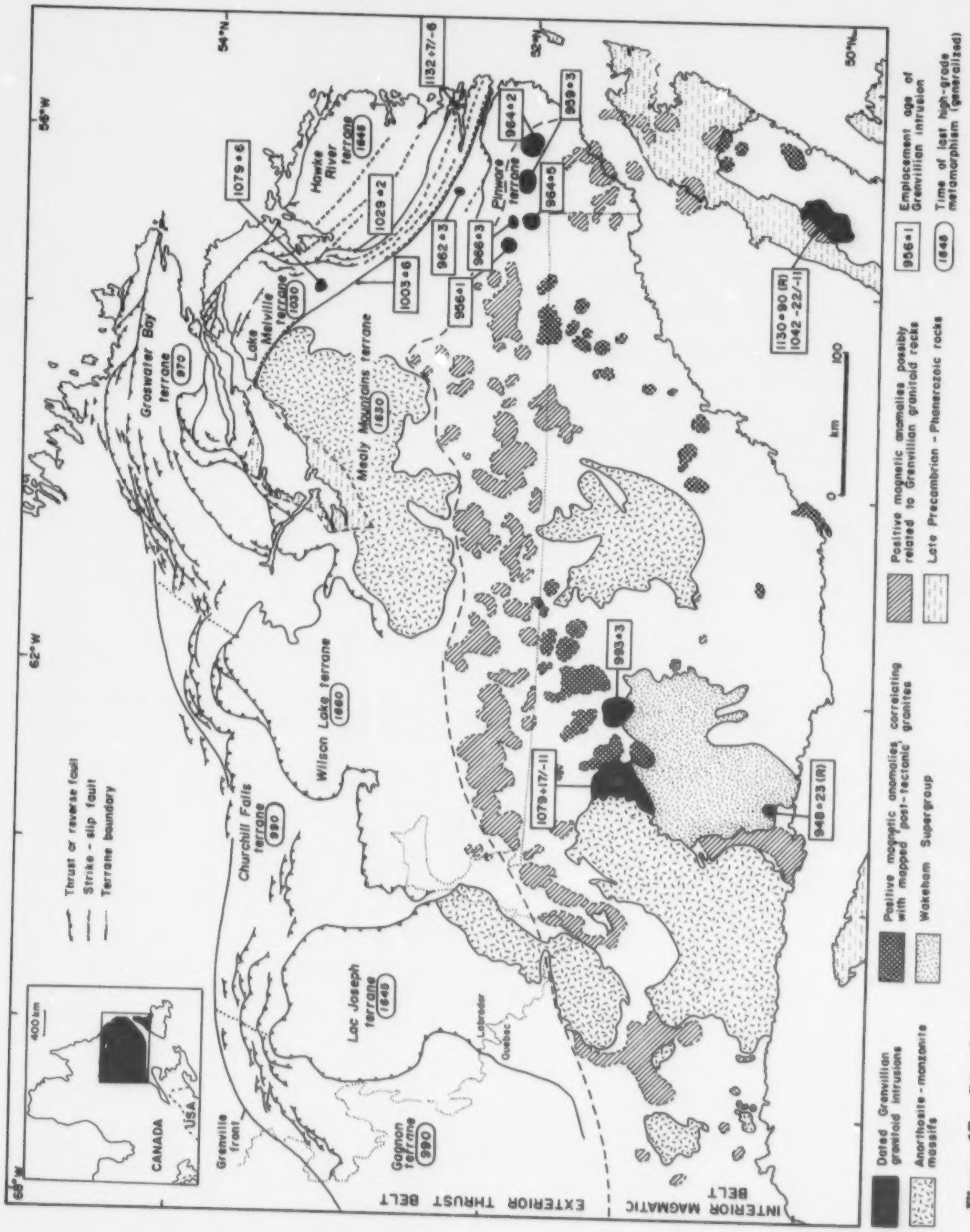
CANADA
USA
400 km
Thrust or reverse fault
Strike - slip fault
Terrane boundary
Grenville front
Gagnon terrane
990
Lac Joseph terrane
1648
Churchill Falls terrane
990
Wilson Lake terrane
1660
Mealy Mountains terrane
1630
Groswater Bay terrane
970
Lake Melville terrane
1030
Hawke River terrane
1648
Pinware terrane
Labrador
Quebec
EXTERIOR THRUST BELT
INTERIOR MAGMATIC BELT
62°W
56°W
54°N
52°N
50°N
1079+6
1029+2
1003+6
982+3
966+3
956+1
964+2
964+5
959+3
1132+7/-6
983+3
1079+17/-11
948+23 (R)
1130+90 (R)
1042 -22/-11
0
100
km
Dated Grenvillian granitoid intrusions
Anorthosite - monzonite massifs
Positive magnetic anomalies correlating with mapped 'post-tectonic' granites
Wakeham Supergroup
Positive magnetic anomalies possibly related to Grenvillian granitoid rocks
Late Precambrian - Phanerozoic rocks
956+1
Emplacement age of Grenvillian intrusion
1648
Time of last high-grade metamorphism (generalized)

Figure 17. *Distribution of Grenvillian granitoid intrusions in the eastern Grenville Province. Modified from Gower et al. (1991).*

Apart from straightforward comparative metallogeny, there are other underlying themes in this article. First, the long history of mineral exploration and geological research in Baltica offers a wealth of information, with which geologists involved in mineral exploration in eastern Laurentia would do well to become familiar. Second, the Scandinavian experience emphasizes that success does not come easily—the fact that economic deposits were not discovered during an earlier exploration phase does not mean they do not exist (the discovery of the Tellnes deposit is a good example). Third, despite exploration extending over decades, or even centuries, new types of deposits are still being discovered, in some cases in regions where mineralization was previously unknown. Finally, it must be stressed that Baltica and Laurentia are not identical. Marginally suitable tectonic environments containing only non-commercial mineralization in one region may well correlate with ideal settings hosting major deposits in the other.

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GEOLOGY OF THE SOUTHERN PART OF THE ARCHEAN ASHUANIPI COMPLEX, SHABOGAMO LAKE MAP SHEET (NTS 23G), WESTERN LABRADOR

D.T. James and D.M. Stephenson[1]
Labrador Mapping Section

ABSTRACT

In western Labrador, the area west of the Proterozoic Labrador Trough is underlain by Archean rocks of the Superior Province. These rocks, which are mainly orthopyroxene ± garnet diatexite, metasedimentary gneiss and metatonalite, and lesser amounts of mafic gneiss and metapyroxenite, make up the southeast corner of a 90,000 km² high-grade gneiss domain named the Ashuanipi Complex. In the study area, field relations indicate that the diatexite intruded the metasedimentary gneiss and metatonalite after these units had already attained peak granulite-facies conditions, as opposed to a model of diatexite petrogenesis involving a progressive migmatization and in situ derivation from the surrounding rocks. Examination of structural, metamorphic and intrusive relations indicates several ages of migmatization and folding, although these events are all inferred to have occurred during one, late Archean tectonothermal event.

Reconnaissance-scale geochemical studies have shown anomalous gold values from lake-sediment and local bedrock samples in the Ashuanipi Complex, and although no specific gold targets were located in 1991, the Ashuanipi Complex has some exploration potential.

INTRODUCTION

PURPOSE

In 1991, a systematic 1:100 000-scale mapping program in the Archean Ashuanipi Complex, which forms the southeastern edge of the exposed Superior Province, was conducted in western Labrador. Mapping occurred in NTS 23G (Figure 1), including NTS 23G/7, 10, 11, 13, 14 and 15. Mapping was accomplished by helicopter-supported traverses and ground traverses; these being mainly conducted from a base camp on the Churchill Falls–Esker road.

Prior to the 1991 field season, this area had only been covered by small-scale, reconnaissance mapping, hence the geology was not known in adequate enough detail to meet current exploration needs. The purpose of the 1991 mapping program was to meet these needs by mapping the distribution of rock units and major structures. The results of this field program will provide a regional geological context for the gold reconnaissance investigation of the Ashuanipi Complex carried out by the Newfoundland Department of Mines and Energy in the 1980's, and earlier, lake-sediment geochemistry programs carried out jointly with the Geological Survey of Canada. Gold mineralization discovered in high-grade supracrustal rocks of the Ashuanipi Complex north of Schefferville, Quebec (Lapointe, 1986), and gold occurrences in other high-grade Archean terranes (e.g., Böhmke and Varndell, 1986; Barnicoat *et al.*, 1991) suggest that this area could have comparable economic potential. This study will also make a contribution to the overall understanding of structural, metamorphic and intrusive relations in this part of the Ashuanipi Complex.

PREVIOUS INVESTIGATIONS

The area has been previously covered by 1":4-mile-scale mapping by Stevenson (1964) and Fahrig (1967). Parts of the area have also been given a cursory examination by Rivers (1985b, c) as part of 1:100 000-scale mapping of the Early Proterozoic Knob Lake Group and younger Proterozoic sedimentary and intrusive rocks, which occur to the east and south of the map area. The area has also been covered by a regional-scale, lake-sediment geochemistry program (Geological Survey of Canada, 1982a, b) and subsequent gold-related studies by Thomas and Butler (1987), Butler (1987), McConnell *et al.* (1987) and McConnell and Newman

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[1] #2-1052 LaSalle Blvd., Sudbury, Ontario P3A 1X9

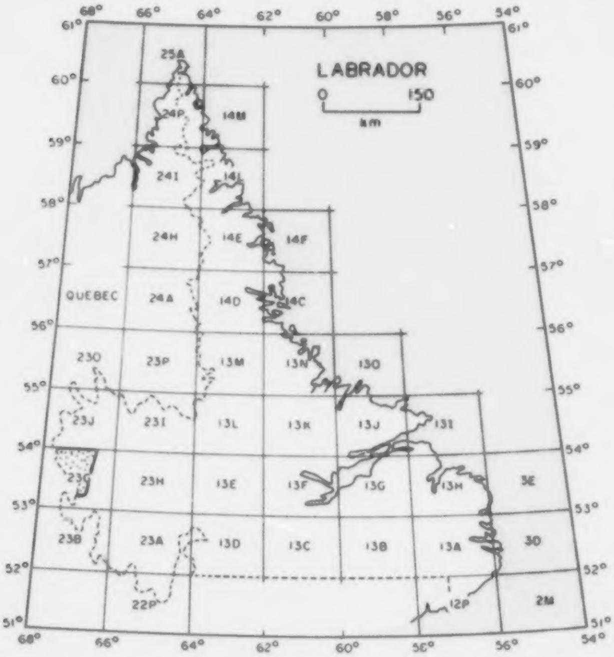

Figure 1. *Index map of Labrador showing the location of the map area (stippled).*

(1988), which concentrated on locating the source of geochemical anomalies discovered during the regional, lake-sediment geochemistry program.

The area mapped in 1991 represents a southward continuation of the 1:125 000-scale mapping of the Ashuanipi Complex in NTS 23J by Percival (1987, 1989) and Percival and Girard (1988). Prior to Percival's work, the Ashuanipi Complex in NTS 23J had been mapped at a scale of 1":4 miles by Frarey (1961) and Stevenson (1963), and farther north, in NTS 23O, by Baragar (1967).

To the southwest of the present study area, parts of the Ashuanipi Complex in Labrador have been mapped by Rivers (1985a).

GENERAL GEOLOGY

REGIONAL SETTING

The Archean rocks mapped in 1991 occur along the southeastern margin of the exposed Archean Superior Province and form part of a 90,000 km² high-grade gneiss domain that is mainly in Quebec (Figure 2). As a result of the initial reconnaissance mapping of this domain in western Labrador in the 1950's and 1960's, which was conducted as part of more detailed investigations of the iron-ore-bearing Lower Proterozoic sedimentary rocks of the Labrador Trough, the Archean metamorphic and intrusive rocks of the Superior Province were collectively named the Ashuanipi Group by Frarey (1961). Subsequently, the rocks were renamed the Ashuanipi Complex by Fahrig (1967). Recently, Card and Ciesielski (1986) have elevated the Ashuanipi Complex to subprovince status, but the term Ashuanipi Complex is still in widespread use (e.g., Percival, 1987, 1989, 1990; 1991a; Percival and Girard, 1988; Machado and Chevé, 1991).

The Ashuanipi Complex is composed of high-grade supracrustal rocks, mainly biotite migmatite derived from greywacke, and metamorphosed plutonic rocks, orthopyroxene-bearing granitoid rocks, and minor, unmetamorphosed tonalite, granodiorite, granite and rare syenite plutons (Eade, 1966; Percival, 1987, 1989, 1990, 1991a, in press; Percival and Girard, 1988; Machado and Chevé, 1991). Most of the complex is composed of granulite-grade rocks, although there are zones of retrogression to amphibolite facies (Herd, 1978).

A sequence of events in the eastern part of the Ashuanipi Complex established by conventional and ion-probe U–Pb geochronology suggests the following Archean history:

1) deposition of sedimentary rocks at approximately 2.7 Ga (Mortensen and Percival, 1987; Chevé and Brouillette, 1991; Percival *et al.*, 1991);

2) intrusion of tonalite plutons and sills into the supracrustal rocks at 2.7–2.69 Ga (Chevé and Brouillette, 1991; Machado and Chevé, 1991; Percival, 1991a; Percival *et al.*, 1991);

3) granulite-facies metamorphism and voluminous, syn- to late syn-metamorphic orthopyroxene-bearing granodiorite (diatexite) intrusions; these are approximately constrained to the interval 2.685 to 2.650 Ga (Mortensen and Percival, 1987; Percival *et al.*, 1988; Percival *et al.*, 1991; Chevé and Brouillette, 1991);

4) initiation of cooling following peak metamorphism at 2.67–2.64 Ga (this is based on U–Pb monazite ages; Mortensen and Percival, 1987; Percival *et al.*, 1988; Percival *et al.*, 1991); the monazite ages suggest rapid cooling following the metamorphic peak (Percival, 1990);

5) intrusion of late pegmatites; one of which is dated by U–Pb monazite to be 2654 ± 5 Ma (Mortensen and Percival; 1987), and leucogranite dated at 2.65 Ga (Percival *et al.*, 1991);

6) post-metamorphic peak thermal events; these are dated by growth of new zircon in a diatexite at 2642 Ma (Mortensen and Percival, 1987), and monazite at 2619 ± 6 Ma (Chevé and Brouillette, 1991; Machado and Chevé, 1991), and

7) intrusion of nepheline syenite at 2625 ±3 Ma (Chevé and Brouillette, 1991; Machado and Chevé, 1991).

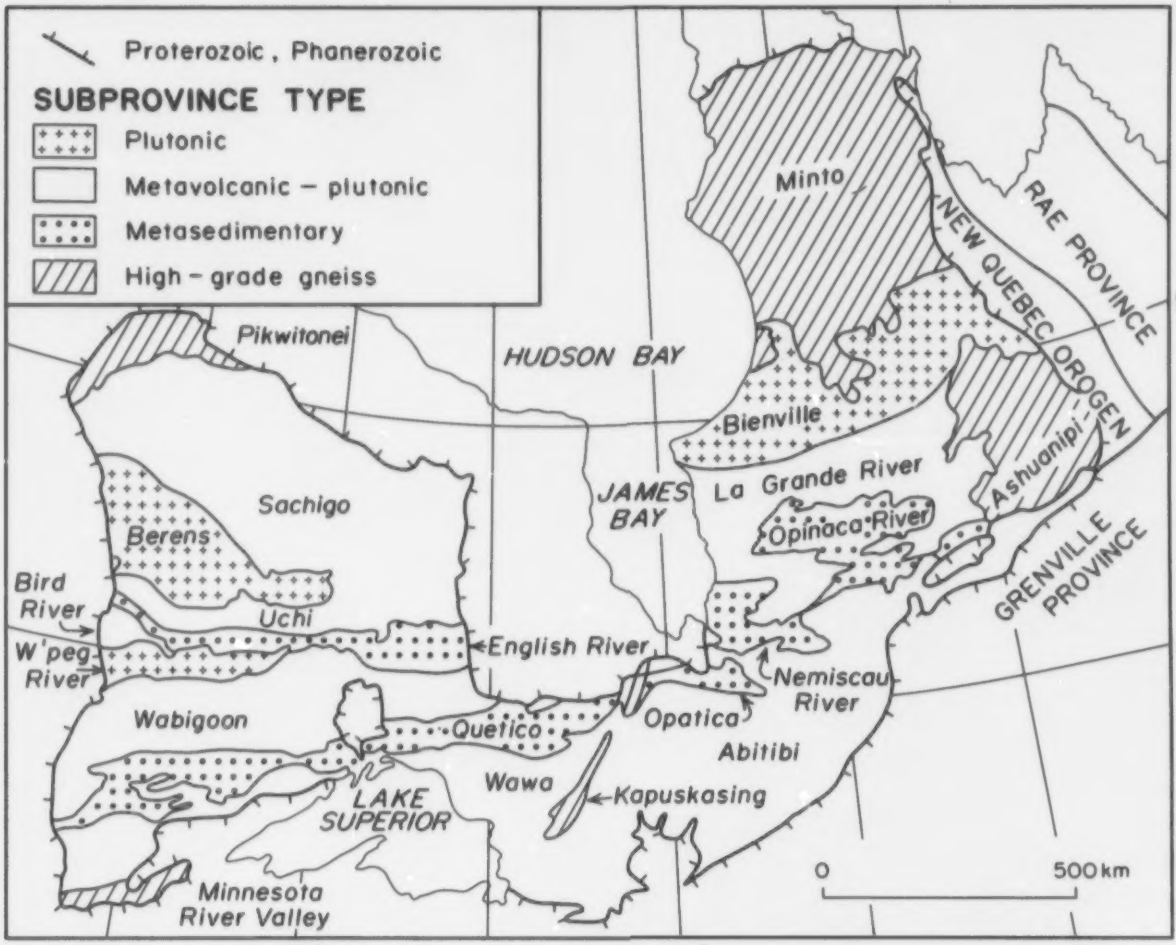
Proterozoic, Phanerozoic
SUBPROVINCE TYPE
Plutonic
Metavolcanic – plutonic
Metasedimentary
High-grade gneiss
Pikwitonei
HUDSON BAY
Minto
RAE PROVINCE
NEW QUEBEC OROGEN
Bienville
La Grande River
JAMES BAY
Opinaca River
Ashuanipi
Sachigo
Berens
Bird River
Uchi
W'peg River
English River
GRENVILLE PROVINCE
Nemiscau River
Opatica
Wabigoon
Quetico
Abitibi
Wawa
Kapuskasing
LAKE SUPERIOR
Minnesota River Valley
0
500 km
55°
50°
90°
80°
70°

Figure 2. *Generalized geological map of the Superior Province showing the distribution of subprovinces (modified after Card and Ciesieksli, 1986). The map area is located in the southeast corner of the Ashuanipi Complex.*

To the west, the Ashuanipi Complex passes into the lower grade Opinaca River and La Grande River subprovinces, which are composed dominantly of metasedimentary rocks, and metamorphosed volcanic and plutonic rocks respectively (Card and Ciesielski, 1986). The boundary between the Ashuanipi Complex, and the Opinaca River and La Grande River subprovinces is a gradational amphibolite–granulite transition zone that has both, prograde and retrograde components (Percival, 1990). To the north, the Ashuanipi Complex is bound by the Bienville Subprovince (Card and Ciesielski, 1986), which is composed mainly of orthogneiss, but has a minor amount of supracrustal rocks (Ciesielski, 1991).

On the larger scale, Mortensen and Percival (1987) have recognized a similarity in rock types, structural style and orientation, and a synchroneity in intrusive and metamorphic events between the Ashuanipi Complex and the Quetico Subprovince of northwestern Ontario. On this basis, the Ashuanipi Complex is considered to be a continuation of the major metasedimentary subprovinces that occur in the central part of the Superior Province (Percival, 1990, *in press;* Percival *et al.*, 1991). From east to west, these subprovinces are the Ashuanipi, Opinaca River, Nemiscau River, Opatica and Quetico; these collectively making up a metasedimentary-dominated belt that is more than 2100 km long. Percival and Williams (1989) and Williams (1990) speculate that the Quetico Subprovince developed as a sedimentary prism between accreting volcanic arcs.

To the east, the Ashuanipi Complex is unconformably overlain by the Early Proterozoic Knob Lake Group (Frarey and Duffell, 1964) of the Labrador Trough. Along the southeastern and southern boundary of the Ashuanipi Complex, the Ashuanipi Complex–Knob Lake Group contact is a Grenvillian-age thrust fault, along which the Knob Lake Group rocks are thrust over the Ashuanipi Complex (see Rivers, 1983, 1985a, c; van Gool *et al.*, 1987, 1988; Brown, 1988; Brown *et al.*, 1991). Locally, the thrusting also involves the Ashuanipi Complex rocks.

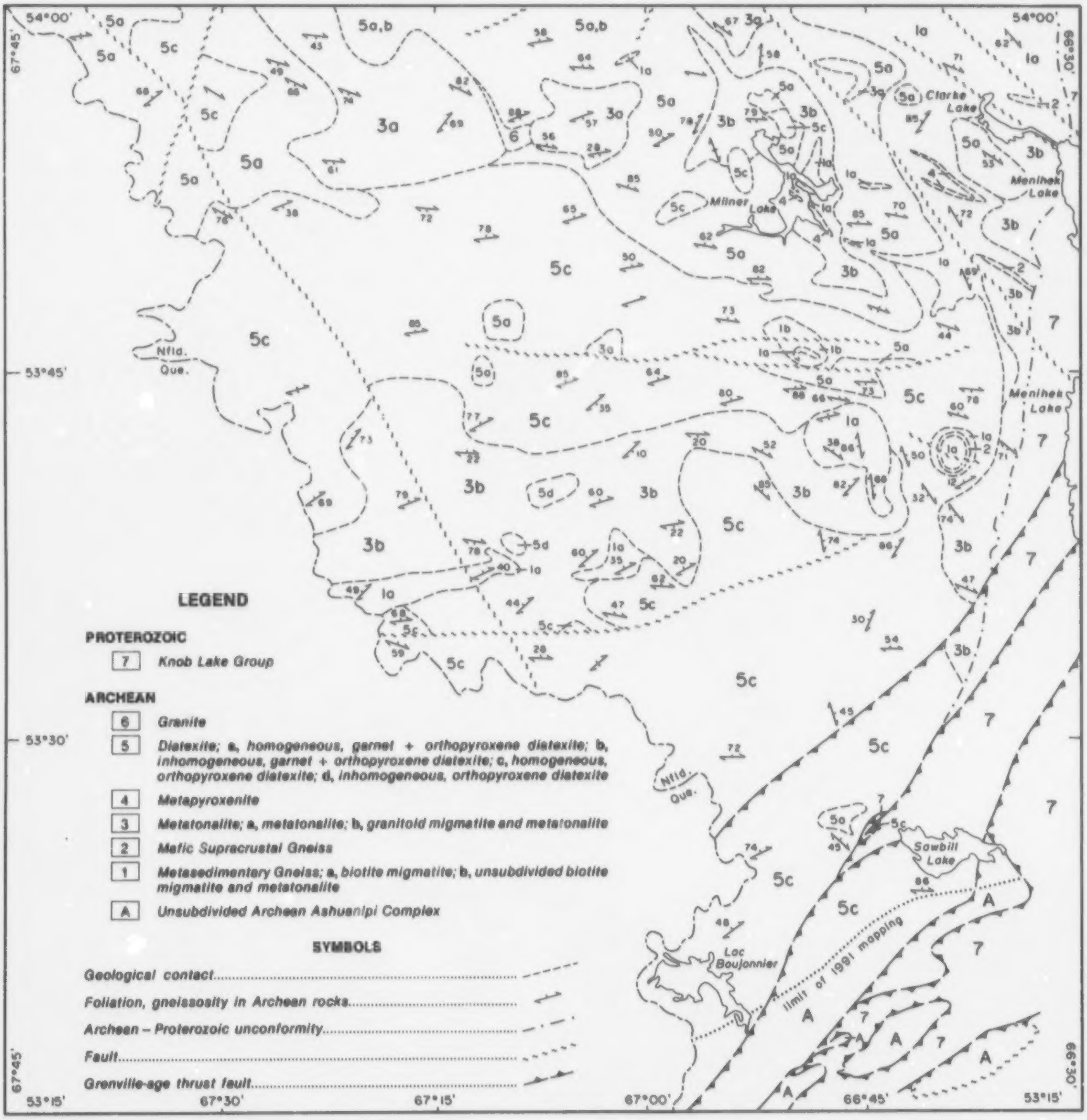
54°00'
53°45'
53°30'
53°15'
67°45'
67°30'
67°15'
67°00'
66°45'
66°30'
Nfld.
Que.
Milner Lake
Clarke Lake
Menihek Lake
Sawbill Lake
Lac Boujonnier
limit of 1991 mapping
LEGEND
PROTEROZOIC
7 Knob Lake Group
ARCHEAN
6 Granite
5 Diatexite; a, homogeneous, garnet + orthopyroxene diatexite; b, inhomogeneous, garnet + orthopyroxene diatexite; c, homogeneous, orthopyroxene diatexite; d, inhomogeneous, orthopyroxene diatexite
4 Metapyroxenite
3 Metatonalite; a, metatonalite; b, granitoid migmatite and metatonalite
2 Mafic Supracrustal Gneiss
1 Metasedimentary Gneiss; a, biotite migmatite; b, unsubdivided biotite migmatite and metatonalite
A Unsubdivided Archean Ashuanipi Complex
SYMBOLS
Geological contact
Foliation, gneissosity in Archean rocks
Archean – Proterozoic unconformity
Fault
Grenville-age thrust fault

Figure 3. *Geological map of the Ashuanipi Complex in part of NTS 23G.*

UNIT DESCRIPTIONS

Metasedimentary Gneiss (Unit 1)

On the basis of intrusive relations observed in the field, the metasedimentary gneiss unit is determined to be the oldest unit in the map area (Figure 3). The unit has been subdivided into two subunits; these are biotite migmatite (subunit 1a), and unsubdivided biotite migmatite and metatonalite (subunit 1b). The biotite migmatite in the two subunits is the same rock type.

The biotite migmatite is composed of grey-, brown- or rusty-weathering restite, and white-weathering leucosome that is present in variable amounts. Locally, the unit consists of greater than 40 percent leucosome. The restite is a fine- to medium-grained, biotite quartzofeldspathic rock that commonly has both garnet and orthopyroxene porphyroblasts (Plate 1). Locally, garnet makes up more than 15 percent of the restite but it is not ubiquitous.

The biotite migmatite has two phases of white-weathering, K-feldspar-bearing leucosome that are

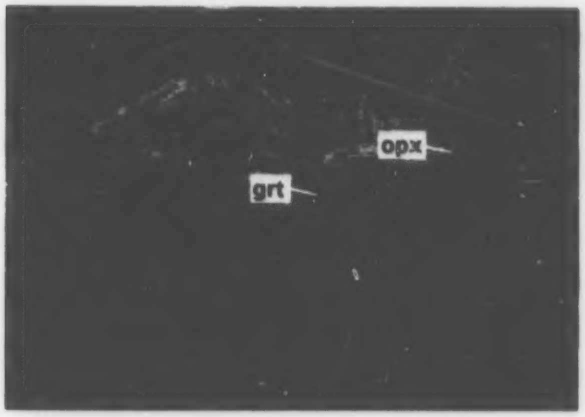

Plate 1. *Metasedimentary gneiss that has garnet (grt) and orthopyroxene (opx) porphyroblasts.*

distinguished on the basis of crosscutting relations and texture. The oldest phase occurs in closely spaced (1 to 10 cm), thin (1 to 2 cm) layers that are oriented parallel to the biotite foliation in the restite; the leucosome imparting a streaky appearance to outcrops. This phase commonly contains garnet and orthopyroxene and makes up less than 15 percent of the unit. The second phase, discordant to both the first leucosome and the biotite foliation in the restite, occurs in layers that are up to 1 m thick and widely spaced relative to the first phase. The second-phase leucosome is fine- to coarse-grained, commonly having coarse-grained feldspars, garnet and orthopyroxene. Outcrops of biotite migmatite may have up to 40 percent of the second-phase leucosome. Layers of second-phase leucosome are deformed.

Locally, the biotite migmatite has a third phase of leucosome that is also distinguished on the basis of crosscutting relations. The third phase is texturally similar to the second-phase leucosome but it does not have garnet or orthopyroxene. The third-phase leucosome is a minor component of the biotite migmatite.

Apart from variations in the amount of leucosome and minor variations in the percentages of garnet and orthopyroxene, the biotite migmatite has a homogeneous composition. These rocks are presumed to be derived from greywacke. Occurrences of mafic boudins, inferred to be derived from pre-metamorphic mafic dykes, are common in the biotite migmatite. Locally, the biotite migmatite unit has thin (1 m), pyrite-bearing zones.

Sheet-like intrusions and small bodies of metatonalite (Unit 3) occur at all scales (up to several hundred metres wide) throughout the biotite migmatite unit. In areas where the metatonalite is particularly abundant and the two rock types cannot be separated at the present scale of mapping, the rocks are mapped as subunit 1b. The contacts between subunits 1a and 1b are gradational.

Mafic Supracrustal Gneiss (Unit 2)

Interlayered locally with the metasedimentary gneiss unit are occurrences of mafic gneiss. Two areas in the northeast corner of NTS 23G/15 have mappable occurrences of these rocks.

The mafic gneiss is black-, black and white-, or rusty-weathering. Outcrops are typically very well layered, consisting of 2- to 5-cm-thick layers that are distinguished on the basis of percentage of mafic minerals, and locally, by thin layers of tonalitic leucosome. The rocks are composed of variable amounts of orthopyroxene, clinopyroxene, hornblende, biotite and plagioclase. Garnet is rare. The leucosome has biotite and orthopyroxene, and makes up less than 15 percent of the unit.

The unit contains local occurrences of sulphide-bearing rocks; these forming 1-m-thick zones that are laterally continuous for several metres in outcrop. Locally, the sulphide zones are strongly magnetic.

The mafic gneiss is presumed to be derived from mafic volcanic rocks. This interpretation is based on the field association with the metasedimentary unit. Relict volcanic features were not observed.

Metatonalite (Unit 3)

The metatonalite unit is subdivided into two subunits designated as 3a and 3b. These are composed of metatonalite (subunit 3a), and a heterogeneous subunit (3b) consisting of granitoid migmatite and metatonalite.

Metatonalite (Subunit 3a)

The metatonalite subunit occurs as sub-mappable, structurally concordant, sheet-like intrusions in the metasedimentary gneiss (Unit 1) and as larger plutons that are surrounded by diatexite (Unit 5). Contacts between the metatonalite and the metasedimentary units are sharp. The metatonalite contains few metasedimentary xenoliths. In contrast, the metatonalite–diatexite contacts are locally gradational, and the diatexite contains abundant metatonalite inclusions in areas adjacent to the contacts.

Metatonalite is structurally and texturally varied. It occurs as variably deformed, metatonalite foliate, which has coarse-grained orthopyroxene and relict igneous textures, and as variably recrystallized, orthopyroxene-bearing metatonalite migmatite.

Most of the occurrences of metatonalite that are within the metasedimentary gneiss unit are metatonalite foliate. These are not shown on the map because of their small size. The metatonalite foliate is white to grey-weathering and fine- to medium-grained having medium- to coarse-grained (to 2 cm), elliptical, inclusion-rich orthopyroxene grains (Plate 2). Based on detailed petrography and mineral chemistry of similar metatonalite from the contiguous Schefferville map sheet (NTS 23J), Percival (1991a) has concluded that the plagioclase- and quartz-inclusion-rich orthopyroxene grains are relict igneous oikocrysts. The rocks also contain minor

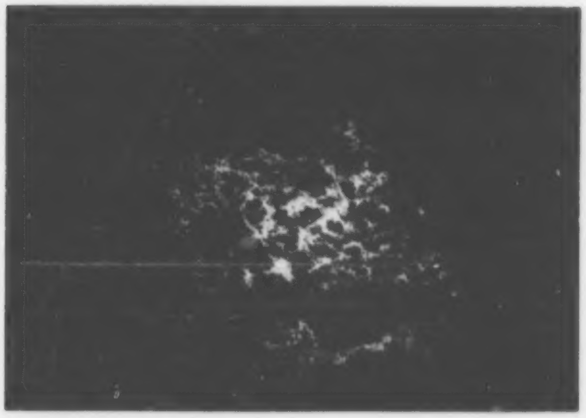

Plate 2. *Typical field aspect of metatonalite (subunit 3a) having coarse-grained orthopyroxene oikocrysts.*

biotite and local garnet. In rare occurrences, the orthopyroxene oikocrysts define layering in the metatonalite that is inferred to be relict igneous layering. The presence of igneous orthopyroxene in these rocks suggests that the tonalite was fractionated from relatively anhydrous magma, possibly in an environment analogous to a Phanerozoic magmatic-arc (Percival, 1991a).

Occurrences of subunit 3a that form the larger metatonalite bodies in the map area are composed of metatonalite foliate, identical to the rocks described above, and as tonalite migmatite. The migmatitic rocks are highly varied in texture and composition. They are variably deformed and recrystallized, and composed of orthopyroxene, common clinopyroxene, plagioclase, biotite, and have up to 30 percent orthopyroxene- and biotite-bearing leucosome. Locally, rocks have two phases of deformed, orthopyroxene-bearing leucosome; these being distinguished on the basis of crosscutting relations. Orthopyroxene occurs in the restite and the leucosome mainly as blocky, equant grains that are inferred to be metamorphic orthopyroxene, and less commonly in the restite as coarse-grained igneous oikocrysts. The rocks contain local garnet. The subunit also contains minor, migmatitic metadiorite and metagabbro. Locally, in the mafic rocks, mafic minerals define compositional layering that may represent relict igneous layering.

The subunit also has a few occurrences of massive to weakly foliated, clinopyroxene-bearing metatonalite. Contact relations between these rocks and the other rocks in subunit 3a were not observed. The clinopyroxene-bearing metatonalite has been correlated with subunit 3a on the basis of rock type, although preservation of a phaneritic texture and the absence of a strong foliation suggests that it may be younger than other Unit 3 rocks.

On the basis of similarity of both rock types and relative age relations, the metatonalite is correlated with the metatonalite that occurs in the contiguous Schefferville map sheet (NTS 23J) mapped by Percival (1989). The U–Pb geochronological studies of the metatonalite in the Schefferville map sheet area indicates igneous crystallization at approximately 2690 Ma; this predates the regional granulite-facies metamorphism by approximately 20 Ma (Percival, 1991a).

Granitoid Migmatite and Metatonalite (Subunit 3b)

Occurrences of granitoid migmatite and metatonalite make up a heterogeneous unit that is designated as subunit 3b. This subunit is distinguished from the metatonalite (subunit 3a) in that it consists of several rock types that have a compositional range from granite to tonalite; the composition being varied down to the metre-scale within subunit 3b. The metatonalite component of subunit 3b is identical to the migmatitic tonalite in subunit 3a, and makes up at least 50 percent of subunit 3b. Contact relations between the different rock types in subunit 3b are equivocal, and they could not be separated at this scale of mapping. Subunit 3b also has common occurrences of leucogranite sheets (Plate 3) and local inclusions of both metasedimentary gneiss (Unit 1) and mafic supracrustal gneiss (Unit 2). The diatextite (Unit 5) has xenoliths of subunit 3b rocks.

Plate 3. *Granitoid migmatite (subunit 3b) having sheet-like intrusions of deformed and recrystallized leucogranite.*

The granitoid migmatite rocks are varied from pink- to grey- to rusty-weathering. They are composed of fine- to medium-grained restite and one or two phases of leucosome, which make up less than 40 percent of the unit. Restite is commonly orthopyroxene-bearing and biotite is ubiquitous. The first phase of leucosome is fine grained and occurs as thin (1 to 5 cm), biotite- or biotite and orthopyroxene-bearing layers. The second phase of leucosome is coarser grained and occurs in layers that are up to 1 m thick; these being discordant to the first phase of leucosome. The second phase of leucosome commonly contains orthopyroxene and biotite, and local magnetite. Both phases of the leucosome are deformed.

Sheet-like intrusions of pink- and white-weathering leucogranite are a common occurrence in the granitoid migmatite subunit; locally, these make up to 20 percent of the subunit. The 1- to 3-m-thick intrusions are discordant to both phases of the leucosome and to the foliation in the restite but are themselves recrystallized and deformed.

Metapyroxenite (Unit 4)

Deformed metapyroxenite dykes and pods occur within both the metasedimentary gneiss (Unit 1) and the metatonalite (Unit 3). They are most common in the northeast corner of NTS 23G/15, where the unit forms dykes in the metasedimentary gneiss. The maximum thickness of a metapyroxenite dyke is postulated to be approximately 250 m although contacts with the surrounding metasedimentary gneiss could not everywhere be precisely located in the field. Local occurrences of small, metapyroxenite inclusions also occur in both the metasedimentary gneiss (Unit 1) and the diatexite (Unit 5) throughout the map area. In the metasedimentary gneiss, the inclusions are interpreted to be relicts of deformed pyroxenite dykes, whereas in the diatexite, the inclusions are inferred to be deformed xenoliths.

The rocks are black-, dark green-, or dung brown-weathering. They are varied from fine to coarse grained, and from massive to foliated. Locally, the rocks are gneissic having fine-scale (less than 1 cm thick) metamorphic layering that is defined by variations in percentage of mafic minerals. The rocks are composed dominantly of coarse-grained clinopyroxene and orthopyroxene, and lesser plagioclase, biotite and possible amphibole. Local olivine is suspected.

These rocks are interpreted to be derived from pre-metamorphic pyroxenite dykes that intruded both the supracrustal rocks and the metatonalite (Unit 3).

The metatonalite (Unit 3) and the metapyroxenite (Unit 4) are correlated with Percival's (1991a) Desliens igneous suite on the basis of similarity of rock types and relative age relations.

Diatexite (Unit 5)

Coarse-grained, orthopyroxene-bearing igneous rocks, which have compositions from granodiorite to monzogranite, are the most abundant rock type in the map area. These rocks are termed diatexites following the definition of Brown (1973), and have been subdivided in the field using the scheme of Percival (1989, *in press*), this being based on the presence or absence of garnet, and the abundance of gneissic inclusions. The subunits are: 1) homogeneous, garnet and orthopyroxene diatexite (subunit 5a), 2) inhomogeneous, garnet and orthopyroxene diatexite (subunit 5b), 3) homogeneous, orthopyroxene diatexite (subunit 5c), and 4) inhomogeneous, orthopyroxene diatexite (subunit 5d). Gneissic enclaves make up to 25 percent of homogeneous diatexite, and from 25 to 50 percent of inhomogeneous diatexite. The gneissic enclaves that are included in diatexite are dominantly metasedimentary gneiss (Unit 1) and metatonalite (Unit 3). Inclusions of mafic gneiss (Unit 2) and metapyroxenite (Unit 4) are less common. The diatexite is intruded by variably deformed, pink- or white-weathering orthopyroxene-bearing pegmatite veins, and minor, leucogranite intrusions (Plate 4).

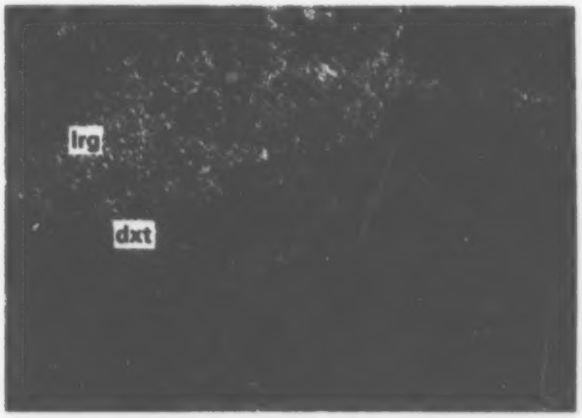

Plate 4. *Typical aspect of homogeneous orthopyroxene ± garnet diatexite (subunits 5a and 5c). Photograph shows diatexite phase (dxt) that surrounds gneissic inclusion and is itself cut by leucogranite phase (lgr).*

The contacts between the diatexite and the older rocks (Units 1 to 4) are narrow zones, at this scale of mapping, but in the field they can be somewhat gradational, consisting of zones up to several-hundred-metres wide, which have older gneiss units and abundant diatexite intrusions. However, the contacts are not transitional contacts that are suggestive of a progressive, *in situ* migmatization of the surrounding rocks. Moreover, the intrusions of diatexite are discordant to the migmatitic structure in the gneissic enclaves (Plate 5), the leucosomes of which are commonly orthopyroxene-bearing; this demonstrates that the surrounding rocks attained granulite-facies conditions prior to the intrusion of the diatexite.

Within the diatexite unit, the quantity of gneissic inclusions is locally quite variable, hence the contacts between the homogeneous and inhomogeneous diatexite subunits are commonly gradational, and the contacts shown on the map are approximate. In contrast, the contacts between the garnet-bearing and garnet-absent homogeneous diatexite appear to be sharp. In the northern part of the map area, where the

Plate 5. *Inclusion of metasedimentary gneiss in diatexite. The gneissic inclusion has an orthopyroxene-bearing metamorphic leucosome that predates the intrusion of the diatexite.*

garnet-bearing variety (subunit 5a) is most common, the garnet-absent diatexite (subunit 5c) occurs in small, discrete bodies within subunit 5a. In the central and southern parts of the map area, the orthopyroxene (garnet-absent) diatexite is dominant; there are only very minor occurrences of diatexite that have garnet.

In individual subunits, there is also local evidence of multiple phases of diatexite that can be distinguished on the basis of grain size and inclusion content. These local, outcrop-scale variations do not form mappable units at the present scale of mapping.

Homogeneous, garnet diatexite (subunit 5a) is white- to rusty-weathering having coarse-grained, K-feldspar phenocrysts and variable amounts of medium- to coarse-grained garnet (up to 15 percent) and blocky, equant orthopyroxene (Plate 6). Locally, both garnet and orthopyroxene have spongy textures that are defined by abundant quartz and feldspar inclusions. Biotite is ubiquitous, and the rocks contain local magnetite. The rocks are variably foliated but commonly the diatexite is massive, the unit having a planar fabric that is defined by flattened, elongate gneissic inclusions (Plate 7). Where the inclusions are abundant, the diatexite has a pseudo-gneissosity having a compositional layering that is defined by the inclusions.

Homogeneous, orthopyroxene diatexite (subunit 5c) is structurally and texturally similar to the garnet diatexite (subunit 5a). Subunit 5c contains up to 15 percent medium- to coarse-grained, blocky orthopyroxene, and lesser biotite and local magnetite. In the central and southern parts of the map area, this rock type has a homogeneous composition and texture over several-hundred square kilometres (Plate 8). The presence of orthopyroxene ± garnet-bearing assemblages in the diatexite indicates granulite-facies, regional metamorphism during formation of the diatexite.

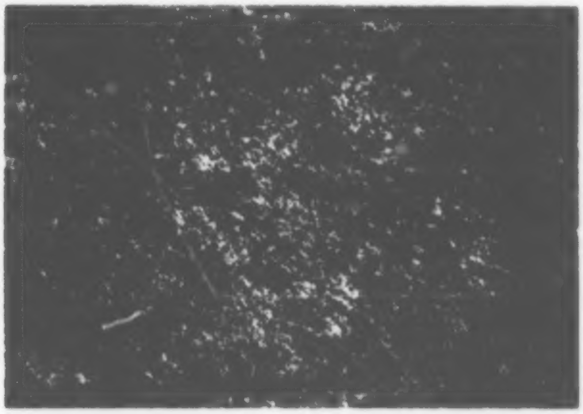

Plate 6. *Typical homogeneous, garnet diatexite (subunit 5a) showing massive structure and coarse-grained texture.*

Plate 7. *Outcrop of diatexite having a foliation that is defined by flattened, elongate inclusions.*

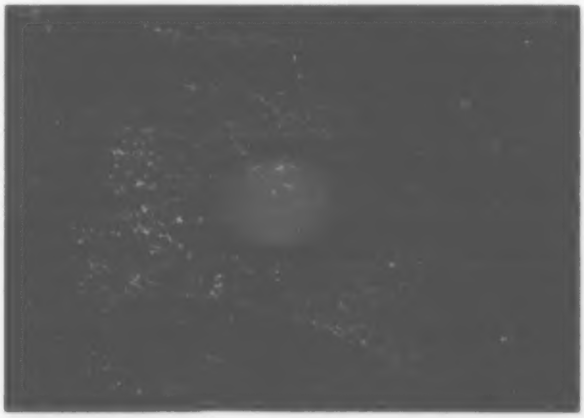

Plate 8. *Typical outcrop of inhomogeneous, orthopyroxene diatexite (subunit 5d) showing coarse-grained orthopyroxene and inclusions of orthopyroxene-bearing metatonalite.*

The non-transitional nature of contacts between the diatexite and the metasedimentary gneiss and metatonalite

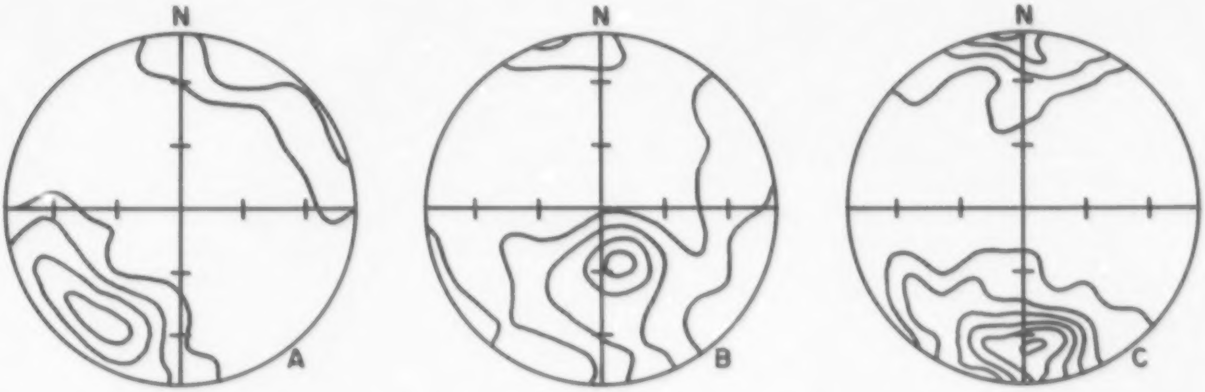

Figure 4. *Contoured, lower-hemisphere, equal area projections of poles to foliation and gneissosity. Contours are in increments of 2σ; A−Poles to foliation and gneissosity (S_1) in metasedimentary gneiss, mafic gneiss, metatonalite and metapyroxenite (Units 1 to 4) in the northern part of the map area (north of 53°45′N). $n=147$, $\sigma=1.41$; B−Poles to foliation and gneissosity (S_1) in metasedimentary gneiss, mafic gneiss, metatonalite and metapyroxenite (Units 1 to 4) in the southern part of the map area (south of 53°45′N). $n=98$, $\sigma=1.37$; C−Poles to foliation in diatexite (Unit 5) in the map area (all diatexite occurrences), $n=264$, $\sigma=1.45$.*

units, and the fact that the diatexite has inclusions which themselves have their own granulite-facies minerals and metamorphic textures, indicate that the diatexite was not derived from *in situ* melting of either the metasedimentary gneiss or the metatonalite. Comparative geochemistry by Percival *(in press)* has shown, however, that the peraluminous, garnet-bearing diatexite has a composition that is identical to the metasedimentary gneiss (i.e., Unit 1), suggesting that melting of metasedimentary gneiss, in a structural level that was lower than the present erosion surface, was the source of the diatexite. Less amounts of melting of the same metasedimentary gneiss source, leaving residual garnet, may account for the garnet-absent diatexite (Percival, *in press*).

The U−Pb zircon dating of diatexite plutons from the Ashuanipi Complex has yielded igneous ages of 2667 ± 1 Ma (Mortensen and Percival, 1987), 2653 ± 3 Ma (Machado and Chevé, 1991), and ages in the range 2685−2670 Ma (Percival *et al.*, 1991).

Granite (Unit 6)

All of the Archean rocks (Units 1 to 5) are intruded by isotropic to weakly foliated and variably recrystallized bodies of leucogranite. Most commonly, these rocks occur as submappable, sheet-like intrusions, although they locally form elongate bodies that can be traced for several kilometres. The larger, map-scale intrusions commonly occur along prominent valleys within the diatexite unit.

The rocks are pink-weathering, typically fine- to medium-grained syenogranite to monzogranite having minor biotite and local magnetite. The massive structure and phaneritic texture suggest that these leucogranite bodies are posttectonic with respect to the main, ca. 2670 Ma metamorphism and deformation in the Ashuanipi Complex.

MAFIC DYKES

The Archean rocks are intruded by three sets of unmetamorphosed and undeformed mafic dykes of presumed Proterozoic age. The dykes are not shown on the map because of their small size. The dykes strike at approximately 055°, 270° and 355°. Occurrences of these rocks are not common in the map area.

STRUCTURE AND METAMORPHISM

A relative chronology of structural, metamorphic and intrusive events can be established for the Archean rocks in the map area based on relations observed in the field. These are summarized in this section.

Local occurrences of compositional layering in the restite of both metasedimentary gneiss and metatonalite units are the oldest structures in the map area. The layering, possibly representing relict bedding and igneous layering, respectively, is completely transposed about the principal foliation (S_1); this being defined by the peak, granulite-facies minerals, orthopyroxene and biotite, in both units. Locally, isoclinal closures of the pre-S_1 compositional layering are observed. In both units, formation of the first phase of leucosome is presumed to be approximately synchronous with the development of S_1 based on the observation that first phase of leucosome is oriented parallel to S_1, and that it contains the peak, granulite-facies assemblage of orthopyroxene, biotite, and local garnet. The principal foliation in the mafic gneiss and metapyroxenite units is also defined by granulite-facies minerals; thus it is correlated with S_1 in the metasedimentary gneiss and metatonalite units. In the northern part of the map area, S_1 in the metasedimentary gneiss, mafic gneiss, metapyroxenite, and metatonalite units is mainly northwest-striking (Figure 4a). In the southern part of the area, the S_1 attitude is varied due to the more pervasive effects of later folding (Figure 4b).

The metasedimentary gneiss commonly contains a second phase of leucosome that is discordant to both the first leucosome and S_1 but also has the peak metamorphic assemblage. In the metasedimentary rocks, the first leucosome and S_1, and to a variable extent, the second leucosome, are folded by tight to isoclinal, northwest-trending folds (Plate 9).

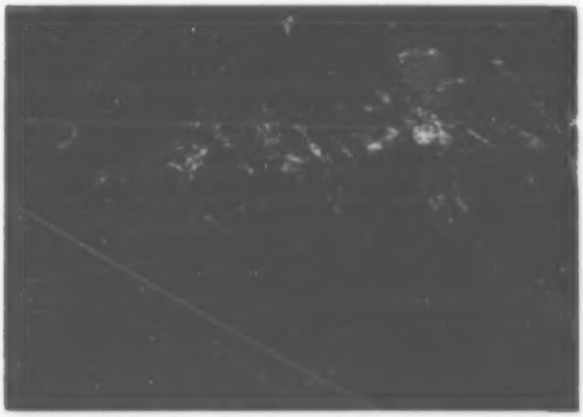

Plate 9. *Northwest-trending fold of S_1 in metasedimentary gneiss that is defined by the biotite foliation and the first phase of the metamorphic leucosome. The fold is transected by lesser deformed second-phase leucosome that is also orthopyroxene-bearing.*

Northwest-trending folds also deform S_1 and the first leucosome in the metatonalite, although the relation between the second leucosome and the folding is unresolved. The second leucosome in the metatonalite is variably deformed and contains orthopyroxene, biotite and local garnet, and has a texture and composition that is similar to the homogeneous, orthopyroxene diatexite (subunit 5c). The northwest-trending folds also deform the metapyroxenite unit.

Intrusion of diatexite postdates both S_1 and the first leucosome in the metasedimentary gneiss, mafic gneiss, metatonalite and metapyroxenite units. This relation is based on observations of inclusions in diatexite that contain both deformed S_1 and the first leucosome. The northwest trend of contacts between diatexite and older rock units in the area suggests that diatexite intrusion may have been approximately late-synchronous with respect to the formation of the northwest-trending structures (i.e., late-syn S_1).

The relation between the diatexite and the second leucosome in the metasedimentary gneiss is unresolved because there were no observed occurrences of metasedimentary gneiss inclusions within the diatexite that contained two phases of leucosome. In contrast, the second leucosome in the metatonalite may be related to the intrusion of diatexite; this based on the aforementioned similarity of rock types and textures.

The diatexite is variably foliated. It has a foliation that is most commonly defined by flattened, elongate inclusions, but also by biotite, and locally by trails of orthopyroxene in the diatexite phase itself. Throughout the map area, the foliation in the diatexite is mainly west-striking (Figure 4c). The discordance in attitude between the west-striking foliation in the diatexite and the northwest-trending structures in the older units suggests that the foliation in the diatexite postdates the northwest-trending folds of S_1 and the first leucosome in the older units. Development of the foliation in the diatexite is inferred to be approximately synchronous with formation of a set of east- to northeast-trending folds that also deform the older rock units and structures. These folds also deform minor intrusions of recrystallized and variably foliated leucogranite that occur mainly within the metatonalite and diatexite units. (These deformed leucogranite intrusions are inferred to predate the undeformed bodies of Unit 6 granite.) The superposition of these folds on the earlier northwest-trending folds has produced map-scale dome and basin structures that are most evident in the southern part of the map area, and has resulted in the preservation of metasedimentary gneiss and metatonalite in structural basins bound by diatexite.

The fact that granulite-facies minerals define an S_1 foliation that predates the intrusion of the diatexite, as well as the foliation in the diatexite itself, and that the deformed leucosomes in the supracrustal rocks and the metatonalite have orthopyroxene and garnet, indicate that the several ages of foliation development, folding and leucosome formation, and intrusion of diatexite occurred during one, granulite-facies event. The U−Pb geochronology by Mortensen and Percival (1987), Percival *et al.* (1988) and Percival *et al.* (1991) of Ashuanipi Complex rocks from north of the map area, approximately constrains the peak of this tectonothermal event to the period 2.69−2.65 Ga.

All of the Archean rock units are deformed by north-northeast-, northwest-, east-, and southeast-striking faults. These faults are consistently marked by prominent topographic lineaments but occurrences of fault rocks are rare. In the map area, the southeast-striking faults are inferred to cut the Ashuanipi Complex−Knob Lake Group contact and have a sinistral transcurrent component of displacement. North of the map area, Percival (1989) has shown that the east-striking faults also cut the Ashuanipi Complex−Knob Lake Group contact. The age of the north-northeast-striking faults is uncertain.

The northeast-striking faults are confined to the southern part of the map area. These faults are Grenville-age thrust faults that extend from the folded and thrusted Knob Lake Group rocks, a pre-Grenvillian autochthonous cover sequence, into the Ashuanipi Complex, the pre-Grenvillian autochthonous basement (see Rivers, 1983, 1985a, c; van Gool *et al.*, 1987, 1988; Brown, 1988; Brown *et al.*, 1991). Exposures of Ashuanipi Complex rocks that are deformed by these faults indicate that faulting occurred under ductile to brittle, greenschist-facies conditions. In the Ashuanipi Complex, the faults are inferred to be steep-dipping reverse faults.

ECONOMIC GEOLOGY

Gold mineralization is not commonly associated with high-grade supracrustal terranes although reported occurrences of gold by Böhmke and Varndell (1986) and Barnicoat *et al.* (1991) from Archean high-grade supracrustal terranes in Zimbabwe and Australia, respectively, suggest that the Ashuanipi Complex may be an exploration target for gold. This is supported by regional geochemical studies in the Ashuanipi Complex by the Geological Survey of Canada (1982a, b) and Hornbrook and Friske (1989), and combined geochemical and geological studies by Thomas and Butler (1987), Butler (1987), McConnell *et al.* (1987) and McConnell and Newman (1988), which showed anomalous concentrations of gold in lake-sediment and local bedrock samples. These results, combined with the reported occurrence of gold associated with metamorphosed iron formation in the Ashuanipi Complex in Quebec by Lapointe (1986), demonstrates that the Ashuanipi Complex has exploration potential.

Mapping in 1991 revealed common occurrences of minor, pyrite-bearing zones within the metasedimentary gneiss and mafic supracrustal gneiss units. Gold analyses of several of these zones yielded results that did not exceed the lower detection limit of 5ppb (approximately). There were no observed occurrences of metamorphosed iron formation in the map area.

SUMMARY

The 1991 field season involved 1:100 000-scale mapping of the Archean Ashuanipi Complex in the northern part of the Shabogamo Lake map sheet (NTS 23G). The area is composed dominantly of orthopyroxene- and garnet-orthopyroxene-bearing diatexite plutons. Field relations suggest that the diatexite is intrusive and not formed as the result of *in situ* migmatization of the surrounding gneisses. The diatexite plutons intrude metasedimentary and mafic gneisses, inferred to be derived from greywacke and mafic volcanic rocks, respectively, and metatonalite and metapyroxenite that are derived from igneous rocks that intruded the supracrustal rocks prior to the high-grade event.

A relative structural, metamorphic and intrusive chronology can be established based on relations observed in the field. The metasedimentary gneiss, mafic gneiss, metatonalite and metapyroxenite units all attained granulite-facies conditions prior to the intrusion of the diatexite plutons. Pre-diatexite structures include the principal, northwest-striking foliation (S_1) in the pre-diatexite rock units, which is defined by both high-grade minerals and metamorphic leucosome, and northwest-trending folds of the foliation and the leucosome. Diatexite, intruded into the surrounding high-grade rocks in bodies that were approximately concordant with the northwest-trending structures (i.e., late syn-S_1) in the supracrustal rocks and metatonalite, have a dominantly west-striking foliation that is defined by both deformed inclusions of the high-grade gneisses and by high-grade minerals in the diatexite itself. The foliation in the diatexite, as well as the older structures, are overprinted by a set of east- to east-northeast-trending folds. The superposition of these folds on the earlier structures has formed dome and basin structures; the supracrustal rocks and metatonalite commonly occurring in structural basins.

Lake-sediment and bedrock geochemical studies by the Geological Survey of Canada (1982a, b), Hornbrook and Friske (1989), Thomas and Butler (1987), and McConnell *et al.* (1987) have shown that there are areas in the Ashuanipi Complex in Labrador of anomalously high gold and gold-sympathetic elements. Mapping in 1991 failed to locate occurrences of bedrock having anomalous gold concentrations; nevertheless, the Ashuanipi Complex has some exploration potential.

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NAIN AREA GEOLOGY: OBSERVATIONS ON SELECTED ISLANDS, AND THE AREA SOUTH OF NAIN BAY (NTS 14C/6, 14; 14D/9)

B. Ryan
Labrador Mapping Section

ABSTRACT

Islands southeast and northeast of Nain surveyed in 1991 are underlain by an amphibolite- to granulite-facies Archean gneiss complex comprising quartzofeldspathic orthogneiss and subordinate amounts of quartzitic supracrustal and mafic gneisses. Deformed and recrystallized noritic and monzonitic rocks akin to those of the Middle Proterozoic Nain Plutonic Suite (NPS) in the area are of unkonwn absolute age. All of the above are intruded by a swarm of variably deformed and recrystallized basic dykes ranging from lamprophyre to gabbronorite (mafic granulite) in composition.

An area of survey south of Nain Bay is underlain by granulite-facies enderbitic to charnockitic gneiss and garnetiferous paragneiss that occupy a northward-tapering wedge between plutonic rocks of the NPS. The NPS of this area has been divided into a gneissic anorthosite-norite, a variably foliated and recrystallized white leuconorite—anorthosite, an undeformed leuconorite, and several elongate, variably deformed intrusions of olivine—clinopyroxene diorite and monzonite. Dykes of the undeformed leuconorite cutting gneissic norite locally contain sulphide mineralization. The deformation of the NPS may be a manifestation of diapiric intrusion of some of its plutons but other explanations are also compatible with the features in these rocks. A discussion of the applicability of a diapiric intrusion model to the plutonic rocks of the Nain Bay area is presented.

INTRODUCTION

A mapping project was undertaken in 1990 to examine a narrow septum of Archean gneisses located between plutons of the Nain Plutonic Suite (NPS) (Ryan, 1991a). As a result of that project, several new aspects of Nain area geology were highlighted. Foremost among these was the recognition of several generations of deformed and metamorphosed basic dykes within the gneisses, and the documentation of strongly deformed and recrystallized rocks within the NPS. The follow-up 1991 project was designed to investigate the areal extent of these phenomena. Gneisses were examined principally on the archipelago northeast and southeast of Nain; a smaller area of gneisses and the enclosing plutonic rocks were surveyed south of Nain Bay (Figure 1). In addition, several helicopter-supported reconnaissance-type sub-projects were undertaken within the NPS to briefly examine selected units of interest shown on unpublished manuscript maps of E.P. Wheeler II, and to collect rock samples for geochronological analyses.

Work on the archipelago was severely hindered during the 1991 season by heavier than normal ice conditions. Access to Nukasusutok Island by boat, to set up the first field camp, was not possible until August 4, and it was six days later before the ocean between there and Seniartlit Islands could be navigated for survey purposes. Rough sea conditions during the latter part of the season precluded satisfactory coverage of the islands east of South Aulatsivik Island.

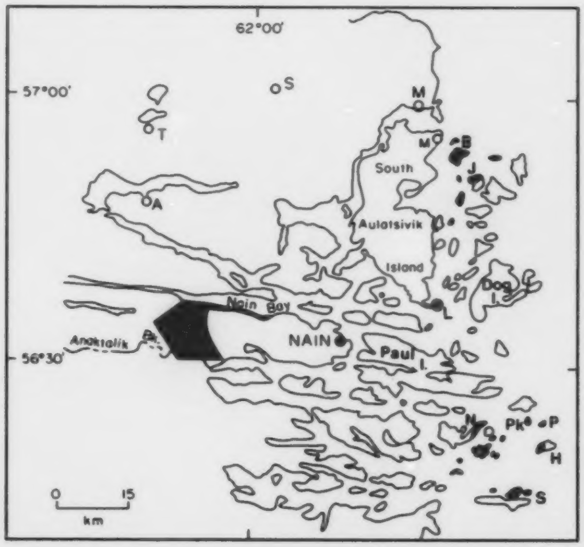


Figure 1. *Map showing the areas surveyed (black) during the 1991 field season. Three areas were examined in detail—between Nain Bay and Anaktalik Brook, south and east of Nukasustok Island (N), and east of South Aulatsivik Island. S = Seniartlit Islands, H = Humbys Island, P = Potuk Island, Pk = Poktoghusekh Island, B = Bouverie Island, J = Jonathan Island. Circles are areas examined briefly and/or rock units collected for geochronological study; L = Loon Island, S = Sheet Hill pluton, M = Manvers granite, T = layered troctolitic body, A = deformed anorthosite—norite.*

GEOLOGY OF THE OFFSHORE ISLANDS

Two areas of Archean gneisses were investigated within the Nain archipelago (Figure 1). These were (i) a cluster of

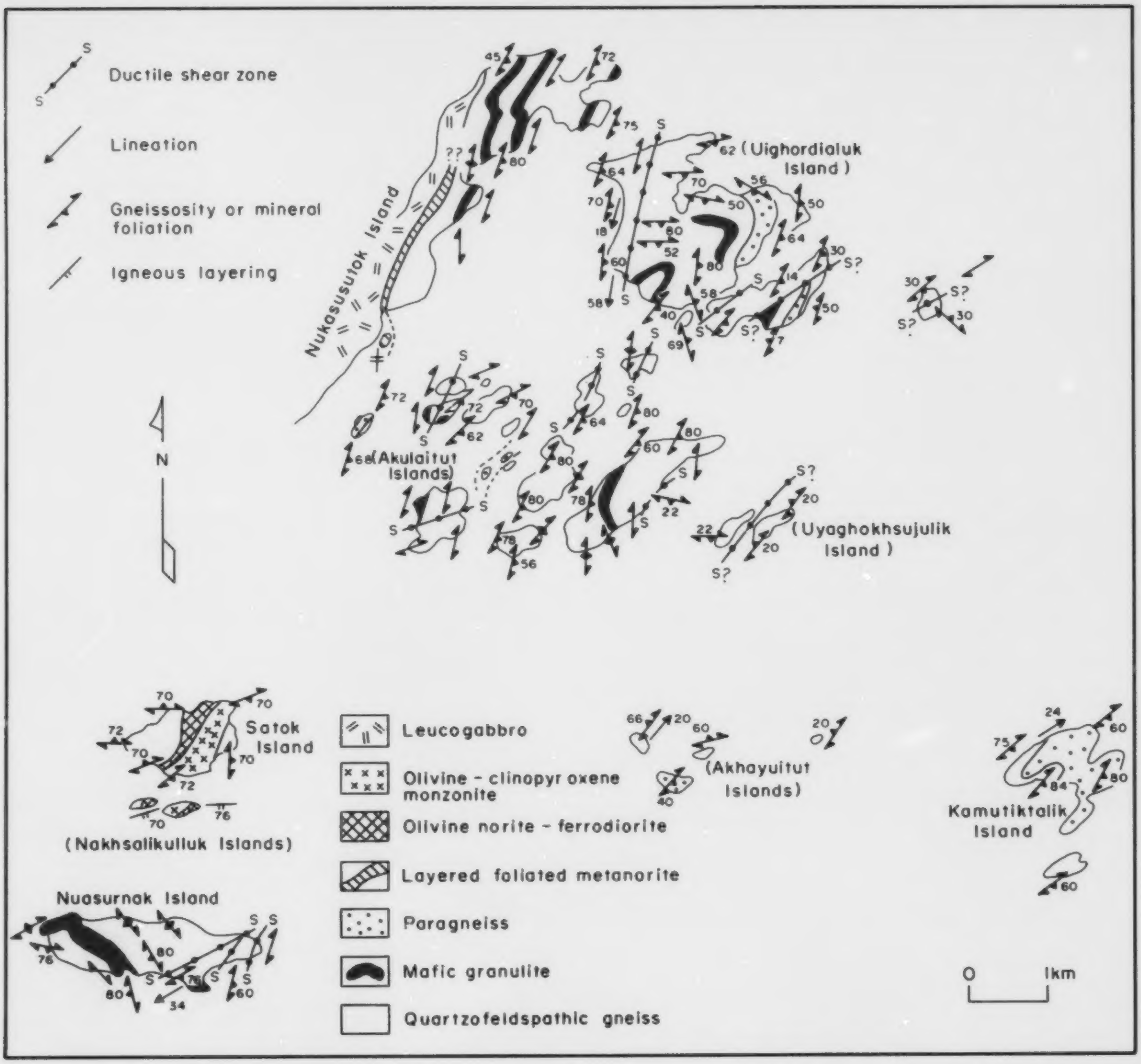

Ductile shear zone
Lineation
Gneissosity or mineral foliation
Igneous layering
N
Nukasusutok Island
(Uighordialuk Island)
(Akulaitut Islands)
(Uyaghokhsujulik Island)
Satok Island
(Nakhsalikulluk Islands)
Nuasurnak Island
(Akhayuitut Islands)
Kamutiktalik Island
Leucogabbro
Olivine - clinopyroxene monzonite
Olivine norite - ferrodiorite
Layered foliated metanorite
Paragneiss
Mafic granulite
Quartzofeldspathic gneiss
0
1km

Figure 2. *Geological map of the area east and south of Nukasusutok Island. Geographic names in parentheses are taken from Wheeler (1953). Names of islands in text that fall outside the boundaries of this map can be found on Figure 1.*

islands east and southeast of Nukasusutok Island and (ii) several islands east of South Aulatsivik (Black) Island.

NUKASUSUTOK ISLAND AREA

Archean gneisses underlie all of the islands directly east of Nukasusutok Island (E.P. Wheeler, manuscript maps; Taylor, 1977; Davies, 1974; Figure 2). Gneisses also occupy the three northernmost islands of the Seniartlit group (Figure 1), the only ones examined. During a brief reconnaissance of Satok Island and small islets directly south of it in 1990, a norite–olivine monzonite sequence of uncertain affinity was discovered; the full extent of these plutonic rocks was determined this year.

The Archean gneisses of the southern project area are dominated by granulite-facies quartzofeldspathic rocks that vary from diffusely layered to distinctly migmatitic. Mafic gneisses are widespread but not abundant; metasedimentary gneisses are also rare. Several generations of granular and schistose mafic dykes intrude all rock types. Ductile shear zones have been identified within this area; these shears postdate some dykes but appear to provide avenues for other (syn-tectonic?) dykes.

Quartzofeldspathic Gneisses

Several types of quartzofeldspathic gneisses have been identified. The majority display granulite-facies mineralogy or are rocks that have retrogressed from the granulite facies. Mesoscopically, these rocks fall into four types: (i) well-layered yellow-brown to grey gneisses with a migmatite component, (ii) schlieric and ghost-layered gneisses, (iii) massive to diffusely layered and agmatitic brown granulite having a pervasive pink granite pegmatite component, and (iv) veined garnetiferous gneisses in which a garnet ± orthopyroxene melt component forms an irregular and sub-parallel network, akin to the Kiyuktok gneisses of Kerr (1980).

Well-layered gneisses (Plate 1) predominate in the central part of the Akulaitut Islands, on Uighordialuk Island and on the Seniartlit Islands. These are variably migmatized, with a *lit-par-lit* and slightly discordant, white to pale pink neosome. Most commonly these rocks are grey to grey brown, reflecting their amphibolite- to granulite-facies metamorphism. Mafic inclusions are abundant throughout these gneisses, but are not of sufficient extent to define units at the 1:50 000 scale of mapping.

Plate 1. *Layered migmatite, Seniartlit Island.*

Schlieric and ghost-layered gneisses have been derived from polyphase granitoid rocks. They contain rafts of older quartzofeldspathic migmatite, mafic gneisses and metasedimentary gneiss, and are the dominant rock type on the northwestern part of the Akulaitut Islands group. A weakly foliated, granulite-facies granitoid, intrusive into mafic granulite units, is the main rock type on the eastern part of Nukasusutok Island and may belong to this same orthogneiss unit.

Brown, rather massive, granulite-facies gneisses underlie the easternmost of the Akulaitut Islands and the southernmost of the central part of the group. Throughout this area, the gneisses contain abundant mafic and ultramafic rafts, and are intruded by a pink pegmatite network of at least two generations. These rocks also outcrop on the two islands south of Uighordialuk Island, on Humby's Island, on Potuk Island and on the northern part of Nuasurnak Island. Retrogressed gneisses, such as those on Poktoghusekh Island, are white weathering.

Kiyuktok-type quartzofeldspathic gneisses (cf. Kerr, 1980; Collerson *et al.*, 1982) are a variation of all the above-mentioned rocks. These gneisses are characterized by garnet ± orthopyroxene-bearing felsic segregations discordant to layering. They are present on parts of the three westernmost islands of the Akulaitut Island group and on the west shore of Poktoghusekh Island. Retrogression of the garnet and orthopyroxene gives these rocks a distinctive blebby texture.

Mafic Gneisses

The mafic gneisses include mafic granulite and amphibolite, the latter being derived by retrogression of the granulite. They occur as small (1 to 10 m^2) podiform bodies or rafts within the quartzofeldspathic gneisses and as units that are hundreds of metres wide and traceable for a kilometre or more along strike. The degree of migmatization is highly variable and individual occurrences may be massive, strike-continuous, banded rocks or they may be pervasively invaded by adjacent gneiss or younger granitoid veining so as to form 'exploded' bodies within a background of quartzofeldspathic material. The most coherent of the mafic gneisses form prominent units on the northeast part of Nukasusutok Island and on west-central Nuasurnak Island.

The mafic gneisses are typically greenish brown or black weathering. They locally have a blotchy texture defined by clots and oikocrysts of black hornblende, and in many outcrops are characterized by irregular veins and 'sweats' of granitic composition that contain conspicuous porphyroblasts of orthopyroxene.

Metasedimentary Gneisses

Gneisses of unequivocal sedimentary parentage are not common in the map-area. The most obvious rocks of this type are quartzites that outcrop on eastern Uighordialuk Island, underlie two small islands of the Akuliatut group, form one island within the Akhayuitut Island group, and appear to be the major rock type on Kamutiktalik Island.

The quartzites are white, grey and buff-coloured, faintly banded, coarsely crystalline rocks; cordierite-, sillimanite- and garnet-bearing pelitic and psammitic units occur with the quartzites on Uighordialuk Island and Akhayuitut Islands. Kamatiktalik Island was not entirely examined, but appears to be largely underlain by grey to brown impure quartzite. These rocks are quartz-rich, biotite-bearing, quartzofeldspathic gneisses that locally contain deep red oikocrysts of garnet. Thin (1 to 3 m) layers of less abundant quartzite, containing thinner (5 to 50 cm) units of orthopyroxene + garnet + cordierite-bearing pelitic gneiss occur at the southern end of the westernmost member of the Akulaitut group, where they are associated with mafic granulite.

Ultramafic Rocks

Orthopyroxenite pods are locally common components within the gneisses, especially in the brown granulite gneisses. They rarely form coherent units at 1:50 000 map scale, but

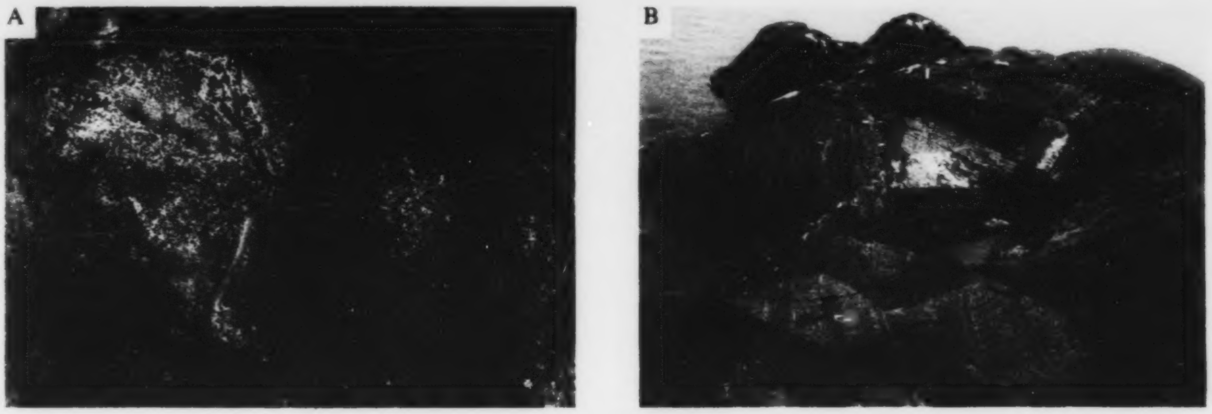


Plate 2. *Metaigneous rocks on Satok and nearby islands. a) Recrystallized noritic blocks in brown monzonite; b) diffuse igneous layering in olivine-bearing noritic phase.*

several of the smaller islands along the northern edge of the Akulaitut group are largely an agmatite of ultramafic blocks within an orthopyroxene-bearing granitoid. The ultramafic rocks are normally pale brown-weathering, but may be dark green or black when clinopyroxene and/or hornblende are present; biotite selvedges occur on such pods in gneisses that have retrogressed. The ultramafic rocks are fine to coarse grained, and vary from massive to well-layered, the latter feature being a relict primary trait of these rocks.

Metaigneous Rocks of Satok Island Area

A short visit to Satok Island during the 1990 field season revealed that the eastern half of the island is dominated by buff- to brown-weathering olivine + clinopyroxene monzonite containing white noritic inclusions (Plate 2a) and cut by deformed hornblende-rich 'amphibolite' and carbonate-bearing lamprophyric dykes (Ryan, 1991a). Further mapping during the 1991 project has defined the extent of these igneous rocks.

The monzonite is a fairly homogeneous, fine- to medium-grained rock and, except for narrow shear zones and a weak foliation, is internally undeformed. Its eastern contact with the gneisses is a mylonite/shear zone, and here the deformation is more severe—the rock has a strong fabric, and lamprophyric dykes are disrupted and drawn out parallel to the contact.

The monzonite is bordered on the west by an oxide-rich, black to rusty brown-weathering, hornblende-rich olivine norite or diorite. The southern contact of this unit with gneisses on Satok Island is a narrow shear zone and the rock is strongly foliated. On the northern part of the island the actual contact with country rock was not observed, but here primary textures, including a diffuse layering, are well preserved. However, the primary character of the mafic rocks is best exhibited in exposures on the Nakhsalikulluk Islands, just south of Satok Island. Here the rock retains many igneous features including a well-developed primary layering in the east (Plate 2b), rafts of leuconorite, rapid textural and grain-size changes, and, in the west, a layered, plagioclase–porphyritic melanocratic gabbroic phase having oikocrysts of black hornblende. The westernmost contact with gneisses, as on Satok Island, is a tectonic one in which a mylonitic structure is present in both rocks. Narrow amphibolitized shear zones also occur in the central part of the unit on these islands.

Metanorite of the Nukasusutok Island area

Wheeler (1969; manuscript maps) defined a 'granulite of uncertain origin' bordering the plutonic rocks on the east side of Nukasusutok Island. Davies (1974) also recognized this unit as being distinct from the surrounding gneisses and referred to it as a 'marginal pyroxene granulite'; she concluded, however, that it is most likely part of the country rock because it is intruded by the Middle Proterozoic igneous rocks of the Nain Plutonic Suite that form the bulk of the island.

The rock is grey to brown weathering and is well layered. The composition is noritic, and in many places on the small island near Nukasusutok Island it still retains vestiges of primary igneous textures (Plate 3a). The mesoscopic layering is on the order of 10 cm to a metre (Plate 3b), and, although enhanced by deformation, is largely an original igneous feature produced by grain size and compositional variations. Some layers are ilmenite and sulphide bearing. Isoclinal folds of the layering are also present, especially in high-strain zones.

The Basic Dykes

Metamorphosed, foliated and folded basic dykes, including rocks with kimberlitic and lamprophyric affinity are present on almost all the islands examined east and south of Nukasusutok. As in the area between Dog Island and Ford Harbour investigated in 1990, several generations are common in many outcrops, and morphological features akin to those described for the Dog Island–Ford Harbour area (Ryan,

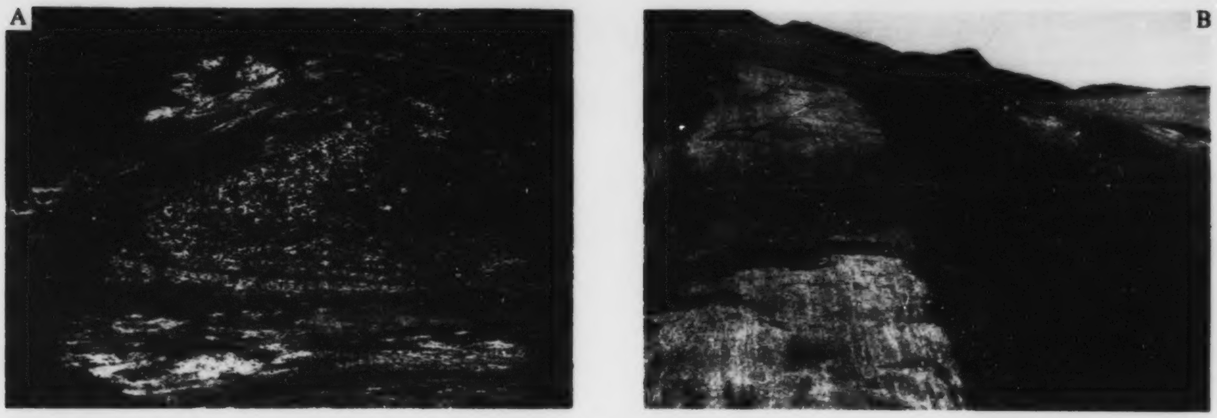


Plate 3. *Deformed, layered metanorite, Nukasusutok Island. a) Inhomogeneous deformation: recrystallized igneous texture in an augen of low deformation surrounded by more highly strained rock; b) compositional layering, reflecting original variations in igneous rock. Darker unit is an ilmenite–sulphide-bearing coarser norite.*

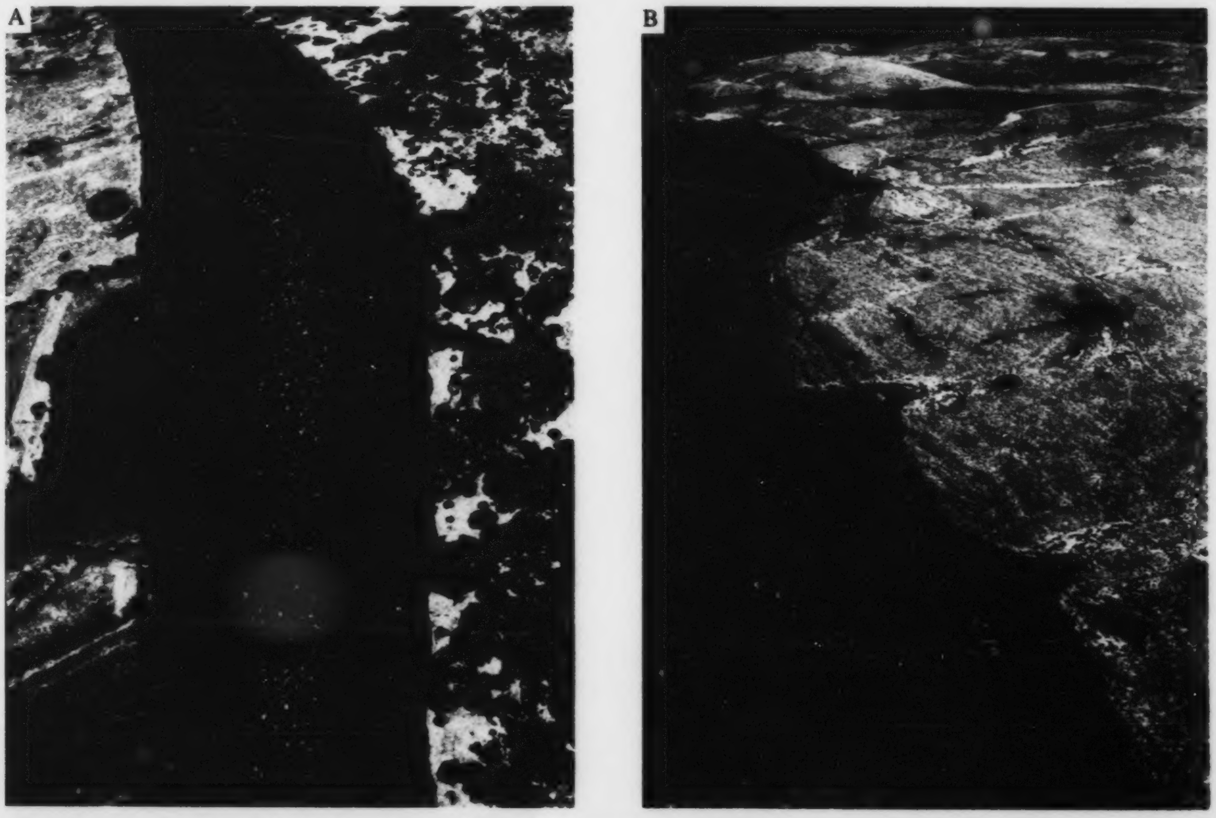


Plate 4. *Metamorphosed basic dykes. a) Basic dyke with central zone of plagioclase concentration and a sigmoidal foliation; symmetry of fabric indicates sinistral shear along plane of dyke. Seniartlit Islands; b) pronounced cuspate margin on basic dyke in grey schlieric gneiss; quartzofeldspathic melts at the tip of the dyke protusions may have been generated at time of dyke intrusion, but similar zones in gneiss distal from the dyke suggest they are earlier features; Akulaitut Island.*

1991a) characterize the dykes observed this year as well. For instance, foliated dykes are common; the foliation is sigmoidal (Plate 4a), oblique and parallel to the walls of the dykes. Folded dykes are less common than foliated ones. Cuspate margins characterize many dykes (Plate 4b), but most are straight-walled. Podiform and pinch-and-swell dykes are also present. Most dykes are steeply dipping, but subhorizontal and moderately dipping dykes occur on the easternmost islands such as the Akhayuitut group, Kamutiktalik Island, and Seniartlit Islands.

The intrusion of some dykes along, and probably concomitant with the formation of, shear zones is clearly demonstrable in many outcrops. It is also apparent that some dykes predate shear-zone development, and in some cases were themselves reactivated as shear zones following intrusion of a crosscutting set of dykes.

The dykes are predominantly black, massive amphibolite and granular mafic granulite. However, a variety that also occurs widely is a greenish brown 'granulite' in which irregular oikocrysts of black hornblende form a random interconnecting network. These hornblende networks produce a discontinuous streaky black fabric in the deformed dykes. Porphyritic dykes, in which plagioclase phenocrysts are concentrated in central or peripheral parts of the dykes, are present locally.

Indications of lamprophyric dykes in the region were noted on southern Satok Island last year (Ryan, 1991a) where black carbonate-rich dykes intrude olivine monzonite. Lamprophyric dykes have now been observed cutting the gneisses on Nuasurnak Island and several of the islands east of Nukasusutok. Most are dense, black, hornblende ± biotite ± carbonate-bearing rocks that exhibit a diffuse compositional layering. They form a sub-parallel group in monzonite along the southeast coast of Satok Island near the contact between the monzonite body and the gneisses. Other dykes of this type are zoned, pitted (weathered ocelli?), greenish black olivine–pyroxene–phlogopite rocks of probable kimberlitic affinity. Lamprophyric dykes are folded and truncated by a northeast-trending shear zone on Nuasurnak Island, plainly indicating that they predate at least one period of zonal ductile deformation in the gneisses.

SOUTH AULATSIVIK (BLACK) ISLAND AREA

The dominant rock type is a brown, diffusely layered to massive quartzofeldspathic granulite, especially well displayed on Little Fish Island and Bulldog Island (Figure 3). Screens and fragments of mafic granulite and ultramafic rock are common. On Little Fish Island and the northern part of Bouverie Island, the gneisses are intruded by numerous northeast-trending dykes of pink granite (probably related to the Manvers granite; Morse, 1969a). A massive grey to pink, diffusely layered granite that outcrops on the western coast of Bouverie Island (Figure 3) may be the same age as these dykes. Grey-brown, well-layered and migmatitic gneisses extend from Turnagain Island northward through Jonathan Island. The interior of Sculpin Island is a rather featureless white granitoid.

An enigmatic unit of layered granulite occurs along the western margin of the Jonathan Island anorthositic-troctolitic intrusion on Jonathan Island and on an adjacent small island. This unit has attributes of mafic rocks adjacent to massive igneous rocks of the NPS on Carey Island and at Higher Bight (Ryan, 1991a). It is a brownish-weathering, equigranular rock characterized in part by an irregular interconnecting network of brown pyroxene veins that, where pervasively developed, give it a brecciated appearance. A fine-grained, equigranular, grey-purple anorthositic to noritic component forms layers within the granulite and both are folded. The anorthositic part is, however, clearly of intrusive origin for it locally randomly transects the more massive pyroxene-veined rock and contains fragments of it. In thicker dykes and sheets it appears to be

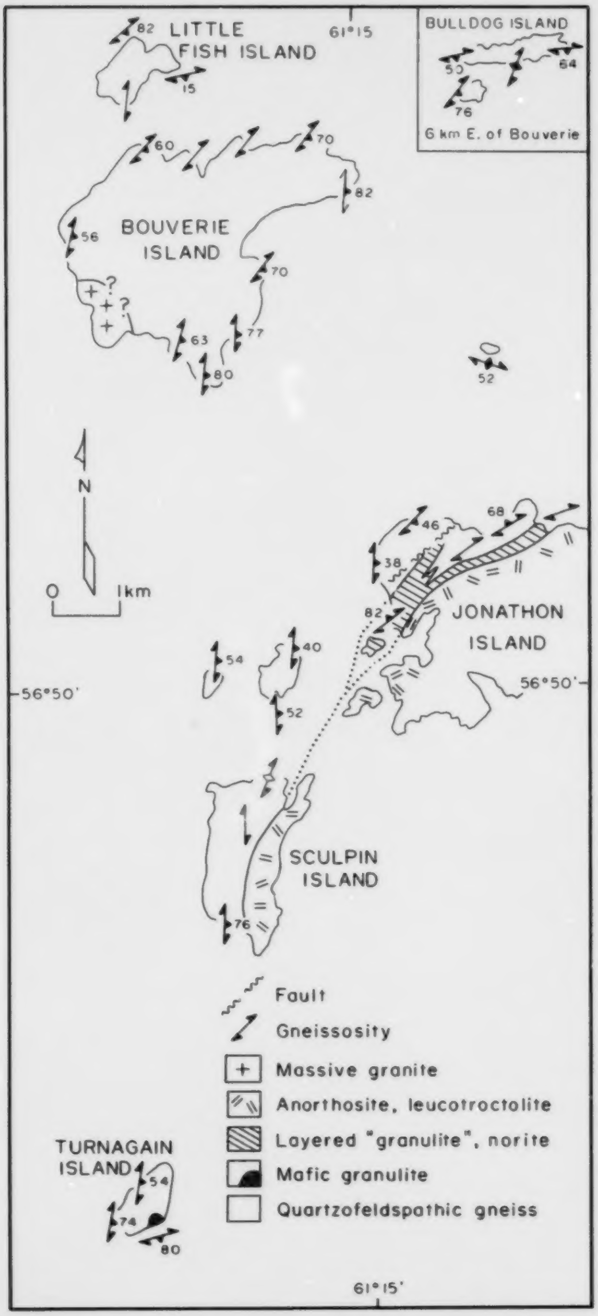


Figure 3. *Geological map of the islands east of Black Island village, South Aulatsivik Island.*

monzonitic, having a white speckled appearance that may denote the presence of potassium feldspar (cf. Berg and Breigel, 1983). On the northeastern shore of Jonathan Island the purplish noritic rock forms the matrix to a breccia that contains fragments of white granular anorthosite (Plate 5). Here the noritic rock also contains sulphides, that give rise to small gossan zones.

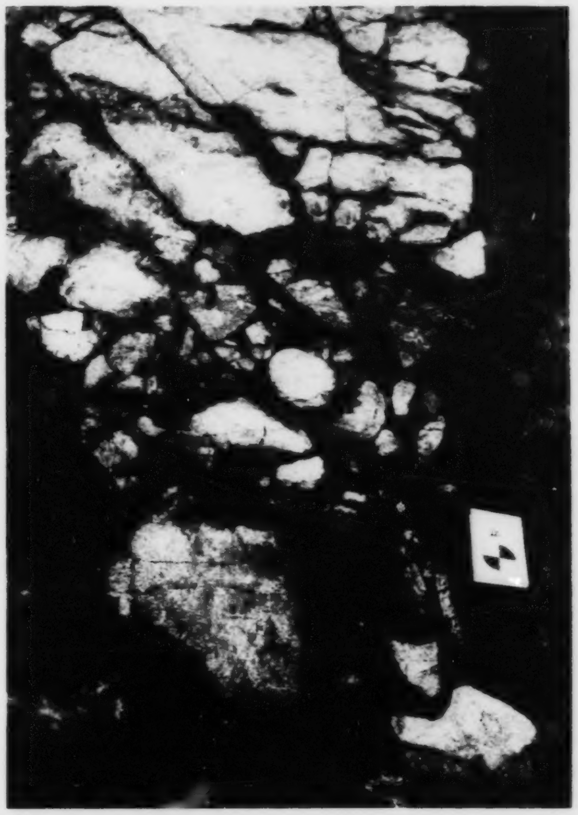


Plate 5. *An intrusion breccia of white fine-grained anorthosite fragments within purple-grey sulphide-bearing norite, northern Jonathan Island.*

Metamorphosed and deformed basic dykes are present throughout the island chain east of Black Island. The south side of Bulldog Island provides a spectacular exposure of such dykes; here east-trending north-dipping dykes within south-dipping, brown, granulite-facies gneisses produce a subhorizontal striped pattern to the nearly vertical 80 m cliff that can be seen from several kilometres distance. Crosscutting relationships in many outcrops point to at least two generations of such dyke intrusions on the islands. Two dykes worthy of note are (i) a curved coarse-grained, 50-m-wide granulite dyke that forms a conspicuous body west of the Jonathan Intrusion on northern Sculpin Island (cf. Berg and Breigel, 1983) and (ii) a foliated 060°-trending granular dyke that contains fragments and globular trains of white norite on the eastern point of Little Fish Island. The former may be related to the Jonathan intrusion, whereas the latter provides more evidence that there are foliated dykes that postdate or are contemporaneous with some of the anorthosites. Several plagioclase porphyritic dykes have been noted, and one of these, an east–west dyke on the eastern headland of Bouverie Island, has a streaky hornblende foliation oblique to its walls and has been dextrally offset, by up to 20 m, along several local north-trending ductile shear zones.

GEOLOGY OF THE AREA SOUTH OF NAIN BAY

A brief reconnaissance of the gneisses in the region between Nain Bay and Anaktalik Brook in 1990 revealed the presence of deformed anorthositic and granitic rocks (Ryan, 1991a). An attempt was made during the 1991 season to further define the extent of the foliated plutonic rocks and also to investigate the nature of the gneisses. It is clear that at least three units of granulite-facies gneiss and several units of anorthositic and dioritic rocks are present (Figure 4).

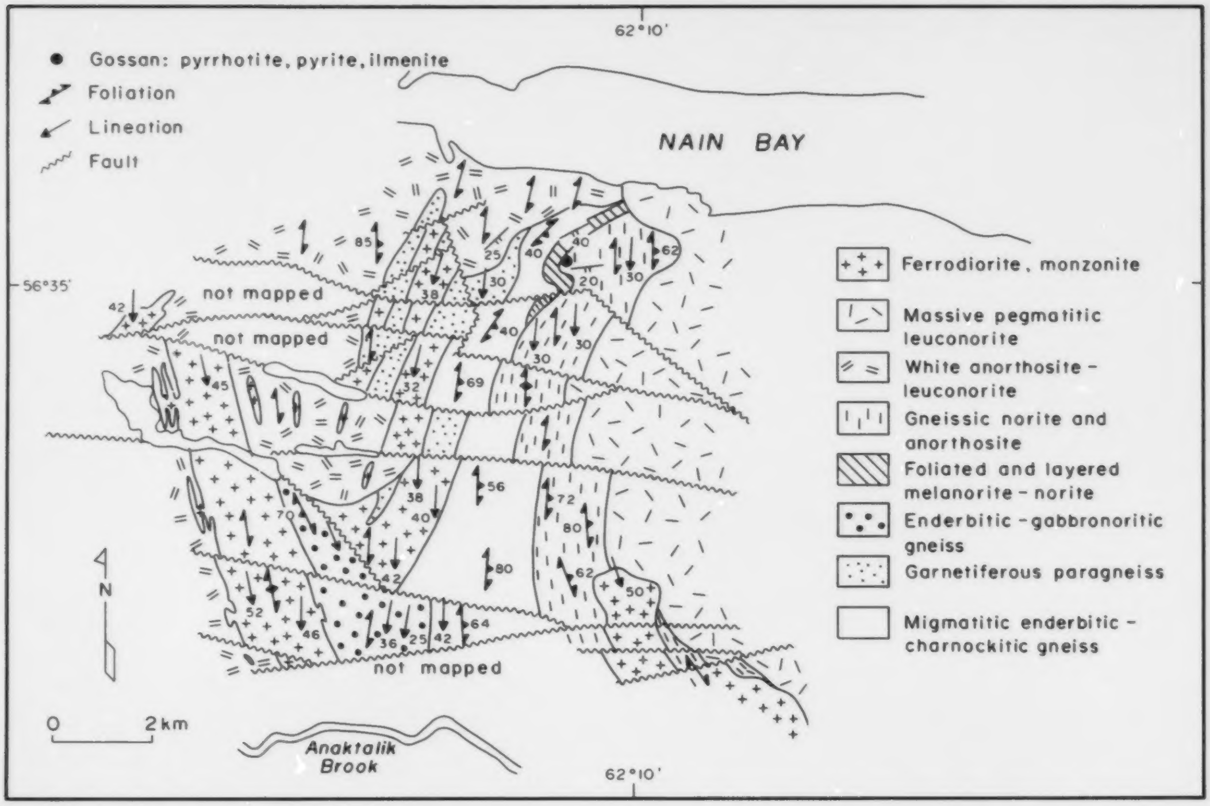
THE GNEISSES

Layered Migmatitic Gneiss

The easternmost unit is a well-layered, somewhat migmatitic, yellow brown-weathering enderbitic and charnockitic gneiss, that locally contains more quartz-poor compositions such as leucocratic mangeritic gneiss, gabbronoritic gneiss and noritic gneiss. In some areas, the gneisses tend to have a more chaotic or agmatitic structure in which mafic pods occur in a rather homogeneous charnockitic host, or charnockitic rocks occur as rafts engulfed by a pink granitoid host. Coherent and podiform mafic granulite units are locally present; these are generally less than 5 m thick and are interpreted to be derived from mafic dykes. Isoclinal folds of the layering are present in some outcrops, but overall the unit is not structurally complicated on a regional scale.

Garnetiferous Gneiss (Tasiuyak Gneiss ?)

The layered charnockitic–enderbitic unit described above is bordered to the west by garnetiferous granitic gneiss and graphitic garnetiferous paragneiss in its northern part, but by a homogeneous, buff-weathering, diffusely layered quartzofeldspathic granulite (see below) to the south. The garnetiferous paragneiss is similar to Tasiuyak gneiss, but differs from it in (i) being garnet-poor and having smaller garnets, (ii) being indistinctly layered, and (iii) being more biotite-rich. In addition the assemblage commonly contains orthopyroxene, a mineral uncommon in typical Tasiuyak gneiss. The occurrence of regional (?) orthopyroxene and the preservation of garnet implies contact metamorphic conditions that differ from the Tasiuyak gneiss within anorthositic plutons elsewhere in this region (cf. Lee, 1987) where garnet is destroyed adjacent to the plutonic rocks. Interbanded with the graphitic and garnetiferous paragneiss are garnetiferous granitoid sheets. These have a *lit-par-lit* morphology, and commonly contain screens of the more biotite-rich metasedimentary rocks.


62°10'
Gossan: pyrrhotite, pyrite, ilmenite
Foliation
Lineation
Fault
NAIN BAY
56°35'
not mapped
not mapped
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0
2 km
Anaktalik Brook
62°10'
Ferrodiorite, monzonite
Massive pegmatitic leuconorite
White anorthosite - leuconorite
Gneissic norite and anorthosite
Foliated and layered melanorite - norite
Enderbitic - gabbronoritic gneiss
Garnetiferous paragneiss
Migmatitic enderbitic - charnockitic gneiss

Figure 4. *Geological map of the gneiss–anorthosite–ferrodiorite–monzonite terrane south of Nain Bay. Some contacts are based on unpublished maps of E.P. Wheeler II.*

Diffusely Layered Enderbitic–Gabbronoritic Gneiss

The third gneiss type in the area between Nain Bay and Anaktalik Brook is a diffusely layered, buff- to orange-brown-weathering, enderbitic to gabbronoritic gneiss. It is not as well-layered as the gneiss of the similar overall composition to the east of it described above, and in places it is simply a foliated igneous rock. It locally contains discrete rafts of older gneiss, and it displays a shallowly south-plunging mineral-streaking lineation. In places it is difficult to distinguish from the foliated, brown-weathering, charnockitic plutonic rocks of the NPS (see below). It resembles, in overall character, an enderbitic to gabbronoritic rock south of Anaktalik Brook that has been dated at ca. 1909^{+33}_{-21} Ma (Krogh and Schärer, 1987).

THE PLUTONIC ROCKS

The plutonic rocks comprise anorthosite–norite and monzonite–diorite types. At least three different units of anorthosite–norite type can be defined, and three elongate diorite–monzonite plutons and numerous dykes of the same are present both in adjacent gneisses and within the anorthosites. The anorthositic rocks display textures that range from strongly foliated and gneissic to massive and subophitic-textured; they are interpreted to include at least three different generations of intrusion. With the exception of deformational foliations and enhanced primary compositional banding in the gneissic unit, oriented structures within the anorthositic rocks are rare. However, an indistinct compositional layering, generally dipping gently to the south can be seen in the cliffs of Nain Bay. The dioritic–monzonitic rocks form three elongate plutons and numerous small dykes in the area. Where emplaced into the gneisses, the larger units are concordant to gneissosity on a regional scale, but slightly transgressive on an outcrop scale. Dykes highly oblique to gneissosity are also present. The two western bodies are locally strongly foliated and exhibit a well-developed lineation, but the southeastern unit is much less deformed. These rocks appear to be all olivine-bearing, and are very conspicuous by their dark to rusty brown-weathering colour.

Gneissic Anorthosite, Leuconorite and Norite

Gneissic anorthosite, leuconorite and norite form a 1200-m-wide unit directly east of the layered gneisses, south of Nain Bay (Figure 4). The unit is compositionally variable over small areas, but a granular white rock of leuconoritic composition prevails. A narrow western border just south of Nain Bay comprises layered brown norite and black hornblende-rich melanorite and lesser ultramafic compositions which, in places, are texturally akin to the metanorite of Nukasusutok Island. The gneissic unit varies mesoscopically from a wispy-foliated rock to a banded gneiss; on hand-specimen scale it is equigranular and thoroughly recrystallized. Some of the feldspar shows rare labradorescence. Very coarse-grained orthopyroxene crystals are present locally, but more commonly they have recrystallized to a granular aggregate, and in most cases are strung-out to give the foliation–banding (Plate 6). Much of the orthopyroxene in the foliated anorthosite–norite is accompanied by, and appears to have been replaced by, hornblende ± biotite. Homogeneous mafic layers ('mafic granulite') occur in parts of the unit, both as strike-continuous or podiform bodies. At several locations in the leuconorite unit, the foliation is crosscut by slightly deformed metamorphosed basic dykes. The southernmost part of this unit is intruded by white to pink aplitic dykes that generally follow the foliation, but are not themselves foliated.

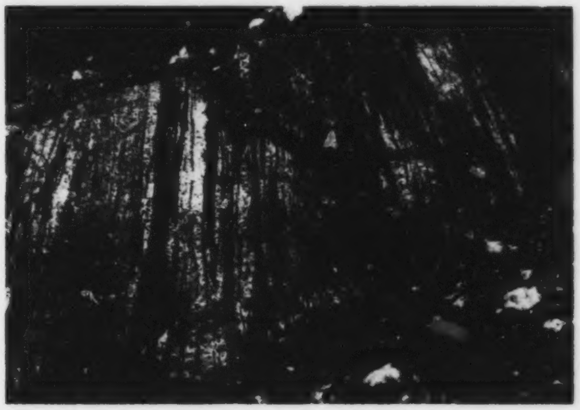

Plate 6. *Discontinuous, streaky aggregates of orthopyroxene, interpreted to be derived from attentuated and recrystallized coarse-grained crystals, south of Nain Bay.*

Foliated Leuconorite and Anorthosite

White-weathering, coarse-grained, ilmenite-bearing, leuconorite and anorthosite, commonly exhibiting a deformational fabric and recrystallized plagioclase, form the westernmost part of the area examined south of Nain Bay. Along its contact, more noritic variants are present, and it is transected by many massive and foliated brown ferrodioritic to monzonitic dykes. The foliation in this rock is not developed to the same extent as it is in the unit above, and is chiefly an augen-type fabric where subophitic pyroxene is drawn into lozenge-like shapes or it is a discontinuous streaky feature defined by flattening of the 'clotted fabric', the latter term coined by Wheeler (1955) for a texture of the anorthositic rocks in which pyroxene is concentrated in irregular zones rather than distributed uniformly through the rock. Directly south of Nain Bay, the foliation is strike-normal to an apparent diffuse layering seen at a distance in steep cliff faces, but where the contact between the gneisses and the southern margin of this pluton was observed, the fabric and compositional layering are parallel to the contact. Internally, this pluton is extremely coarse grained, having kinked crystals of orthopyroxene up to a metre in size; this pluton is probably the same coarse igneous rock that continues westward and hosts the spectacular coarse labradorite of the Pearly Gates area (cf. Watson, 1980).

Pegmatoidal Massive Leuconorite

The easternmost part of the area investigated south of Nain Bay comprises an undeformed, white to grey, coarse-grained leuconorite. This pluton has a well-exposed intrusive contact with the foliated and gneissic anorthosite to metanorite unit described above, and is considered to be the youngest of the anorthositic rocks in this area. The contact with the gneissic norite–anorthosite can be seen on the plateau south of Nain Bay, where a zone, 800 m wide, marks a transition between foliated rock with a stockwork of discordant pegmatitic ilmenite-bearing leuconorite (Plate 7), to leuconorite in which the older rock forms rafts. Where this younger unit has been examined in the north, it is a regionally homogeneous pegmatoidal ilmenite-bearing leuconorite, but a diffuse subhorizontal layering is visible in the 500 m cliffs that form the southern shoreline of Nain Bay. In the south, it is a seriate-textured rock that may in fact belong to an entirely different intrusion.

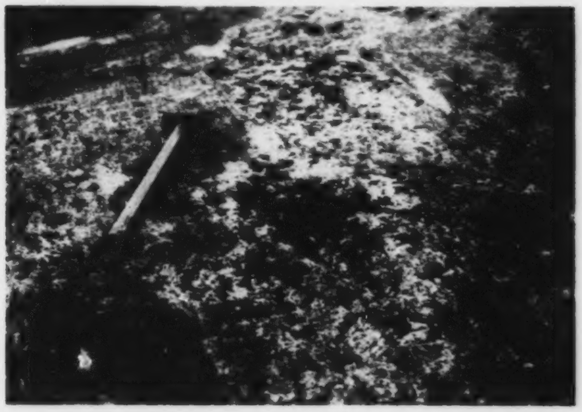

Plate 7. *Recrystallized metanorite having a streaky foliation, intruded by a pegmatoidal dyke of leuconorite; dark area in dyke is a coarse orthopyroxene crystal, south of Nain Bay.*

Western Diorite–Monzonite

The westernmost diorite–monzonite unit is completely enclosed by white weathering anorthosite in the north, but separates anorthosite from charnockitic gneiss in the south. It appears to encompass compositions from olivine meladiorite to olivine–clinopyroxene monzonite (akin to that on Satok Island) and varies from fine grained to coarse grained. Both varieties can be seen in single dykes in many parts of the area. Its western margin is marked by a subvertical sheeted complex of anorthosite and brown monzonite–diorite (Plate 8). Numerous dykes of the monzonitic rock occur throughout the region; the dykes, like the large body, are variably foliated. Inclusions of foliated and recrystallized, white-weathering anorthosite and norite, and of the country-rock gneisses are present in many parts of the intrusion.

Plate 8. *Sheeted contact between anorthosite (white) and ferrodiorite–monzonite comprising approximately 50 percent older and younger rock; note the lensing-out and boudinage of the anorthosite zones.*

Central Diorite–Monzonite

The central dioritic–monzonitic unit is similar to the above, but tends to be overall coarser grained and is more noticably quartz bearing. Unlike the western unit, the central diorite–monzonite is confined largely to the gneisses, and, as noted earlier, is disposed as dyke-like bodies parallel to the regional structural grain. It locally has a megacrystic texture, with some of the perthitic megacrysts having a narrow plagioclase mantle. In this respect, this intrusion is like the undeformed olivine- and clinopyroxene-bearing rapakivi granites that form larger masses within the NPS (cf. Ryan, 1991b). The deformation of the central monzonitic unit is very strong in places (Plate 9). In the originally megacrystic rocks, it has produced an augen structure in which dark streaks of olivine and clinopyroxene are wrapped around the feldspar megacrysts. The foliation is of L > S type in most outcrops.

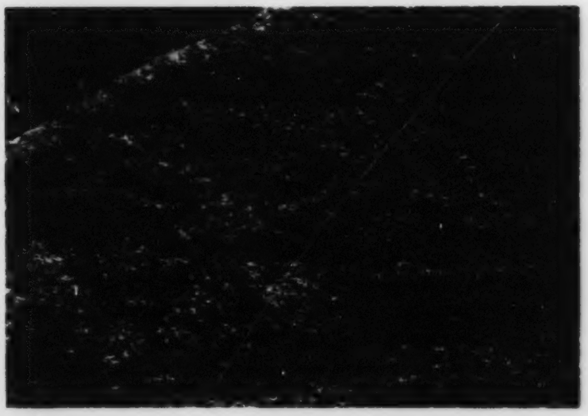

Plate 9. *Foliation in central olivine-bearing diorite–monzonite unit south of Nain Bay; phenocrysts of potassium feldspar stand out in relief.*

Southeastern Diorite–Monzonite

The southeastern body is the least deformed of the dioritic–monzonitic intrusions, although it does display a weak foliation and a south-plunging lineation; the foliation is best defined by the (primary?) orientation of the long axes of tabular potassium feldspar rather than the streaks of mafic mineral aggregates that form the fabric in the two units to the west. In addition to K-feldspar, clinopyroxene and quartz, outcrops also show dark-grey plagioclase xenocrysts in places. Some euhedral to ovoid potassium feldspar phenocrysts have a narrow plagioclase rim. At its southern end this intrusion becomes more granitic, and is characterized by light grey 'drop' quartz, and has abundant gneissic inclusions.

Other Areas Visited

A couple of days at the beginning of the 1991 field season were devoted to helicopter-supported field checking of geologically interesting areas highlighted from an examination of E.P. Wheeler's unpublished manuscript maps, and to collecting samples for geochronological study from difficult-to-reach parts of the NPS. These brief surveys (Figure 1) were carried out in collaboration with R.F. Emslie of the Geological Survey of Canada. In addition, a day was spent with J. Hodych of the Department of Earth Sciences, Memorial University, who collected specimens of mafic dykes for paleomagnetic study.

Among the areas visited in the interior of the NPS is an area 10 km north of Kingurutik Lake (Figure 1, T) shown on E.P. Wheeler's manuscript maps as 'granulite of uncertain origin'. A brief examination indicates this to be a well-layered melatroctolite (or olivine melagabbro) to leucogabbronoritic body having compositional variations on the order of a few centimetres to a metre. The unit is about 1 km in maximum width and is traceable along strike for at least 6 km. Its western boundary with brownish-weathering leuconorite was not investigated, but on its eastern side it is intruded by a massive dark grey to light grey leuconorite to norite.

Wheeler's unpublished maps also show a zone of 'granulite of uncertain origin' stretching several kilometres southward from the central south shore of Kingurutik Lake. A brief visit to one outcrop near the lake's shoreline (Figure 1, A) indicates this unit is foliated and recrystallized, layered, yellow- to white-weathering norite and leuconorite and brown, layered olivine-bearing melagabbronorite akin to rocks that occur on a small island west of Barth Island (Ryan, 1991a). The full extent of this assemblage is unknown since Wheeler's mapping is of a reconnaissance nature in this area, and the writer has not attempted to define the distribution of the deformed rocks outside the outcrop examined. It is likely, however, that gneissic zones of this kind occur elsewhere in the interior of NPS.

The Sheet Hill granite pluton (Figure 1, S; Berg, 1973) was briefly examined, and specimens were collected for zircon geochronological study. This pink-weathering, massive, fluorite-bearing potassic granite was described by

Berg (1973) as being intrusive into the nearby Hettasch layered basic intrusion, although he noted that it was locally 'strongly foliated' and appeared to be in 'gradational contact' with adjacent gneisses. This writer's brief visit to the Sheet Hill granite, suggests the pluton may be older than Hettasch intrusion. The massive interior part of the granite is somewhat misleading, since the conversion of basic dykes within it to hornblende schists implies that a penetrative deformation has affected the granite, and was preferentially absorbed by the dykes. Since the Hettasch intrusion is not deformed, the implication from the deformed dykes is that the Sheet Hill granite is older than the layered basic intrusion. Possible correlations are the Early Proterozoic granites recently identified by Emslie and Loveridge (1991; *in press*) in the Okak Bay area to the north, although the low-magnetic signature of the Sheet Hill granite (G. Kilfoil, personal communication, 1991) sets it apart from the Okak granites. It is possible that the nearby Alliger Lake granite (see Ryan, 1990) is also of Early Proterozoic age.

The Manvers granite, which intrudes the Kiglapait intrusion, was collected at two localities for geochronological study (Figure 1, M), one just west of Village Bay on the shore of Port Manvers Run, the other south of Port Manvers Run at Medusa Bluff (see Plate 1 in Morse, 1969a). The Manvers granite is the name used by Morse (1969a) for a series of dykes within the Kiglapait intrusion that vary from pink aplitic to white pegmatitic rocks; the coarser members contain topaz, fluorite, beryl and amazonite. A sample of homogeneous, pink, medium- to fine-grained granite and a sample of a white pegmatite dyke were collected at Medusa Bluff. Pink granite having zoned plagioclase phenocrysts and opaline ovoid quartz, and a grey, aplitic, speckled biotite granite were collected west of Village Bay. Previous Rb−Sr and K−Ar investigations by Barton (1974) suggest intrusion ca. 1275 Ma, but this is probably a minimum age.

Foliated pink, potassic, hornblende granite and grey-pink granodiorite form the major rock type on the western half of Loon Island (Figure 1, L; cf. Barton, 1977). The granitic rocks are intruded by recrystallized and foliated basic dykes having a variety of orientations, but crosscutting contacts between dykes were not observed during a brief examination of the island. The Loon Island granite resembles granites that intrude mafic granulite and other gneisses on the Red Islands (Ryan, 1991a). All the granites are cut by deformed dykes, and all are of unknown absolute age. A sample suite was collected for U−Pb zircon dating, to ascertain if Barton's (1977) Rb−Sr data indicating a Middle Proterozoic emplacement age were a valid signature.

Several of the metamorphosed and deformed mafic dykes that cut the gneisses were collected during the 1991 field season for geochronological study (see Connelly and Dunning, 1990). Sample sites on Misfit and Dumbell islands (Ryan, 1991a) were revisited during the 1991 field season, and samples of dykes were taken by J. Hodych for paleomagnetic study. Results from these dykes could augment the field and geochronological interpretations and allow constraints to be placed on the time(s) of dyke emplacement and/or metamorphism. At present there are few data on the age(s) of the dykes. Barton (1973) reported preliminary Rb−Sr and K−Ar results that indicated the possibility of three distinct intrusive episodes, namely 1200, 1380 and 1615 Ma. Wiebe (1985) conducted K−Ar whole-rock analyses of two fresh dykes cutting the NPS that indicated minimum ages of 1042 ± 54 and 977 ± 44 Ma.

STRUCTURE AND METAMORPHISM

The overall regional structure of the gneisses of the area is fairly simple. A few large folds can be identified, but no complex patterns are apparent at the mapping scale, although in some outcrops there are clear mesoscopic interference patterns resulting from superimposition of two or more fold generations.

Unique to the Nukasusutok Island area are discrete, ductile high-strain zones in the gneisses (Figure 2). In some cases, these are expressed as layer-parallel zones of abnormal strain within otherwise migmatitic and layered gneisses, but in many instances are highly discordant zones both at outcrop and map scale that juxtapose gneisses with differing orientations. These zones vary from a few centimetres to hundreds of metres in width, and in some cases are occupied by younger, or perhaps contemporaneous, mafic dykes (Plate 10). The post-shear or synkinematic age assignment to these dykes is based on the observation that in some shear zones the dykes are straight-walled and show no penetrative fabric, yet in others the dykes have pinch-and-swell form and exhibit a sigmoidal, oblique or dyke-parallel foliation. It is also apparent, however, that some dykes predate shear-zone formation, for there are dykes that are truncated or drawn into parallelism with the shear zone. Perhaps the best large-scale example of this shear zone−dyke relationship occurs in a cove on the south shore of Nuasurnak Island where lamprophyric dykes in the gneisses are reoriented and folded with their host rocks upon entering a northeast-trending shear zone (Plate 11). Relations between dykes and gneissic layering within the zone indicate initial transposition and rotation of dykes into parallelism with the gneissosity, followed by folding of both as the shear evolved. The shear zones that were examined appear to record two ages of movement at different crustal levels. Some zones retain the high-grade minerals of the adjacent gneisses (i.e., upper amphibolite to granulite facies), whereas others are chloritized, epidotized, and hematized and reflect much lower-grade metamorphic conditions at time of formation. Both types of assemblages can be seen in individual zones, implying the high-grade shears were re-activated at lower P-T conditions during crustal uplift.

Metamorphic assemblages in the gneisses record upper-amphibolite to granulite-facies conditions. The granulite-facies rocks are much more widespread in both the Nukasusutok Island−Humbys Island area and the Bouverie Island−Bulldog Island area than they are in the Dog Island area that was mapped last year. The presence of cordierite as a pseudomorph after sillimanite and garnet in pelitic rocks, by comparison with this transformation elsewhere in the region,

Plate 10. *Sinistral shear zone in gneisses, Akulaitut Islands; three foliated mafic (granulite) dykes occur within the 'straight gneisses' of the shear zone.*

Plate 11. *Oblique aerial view of the margin of a shear zone on Nuasurnak Island; note that lamprophyric dykes on left hand side of photo are folded and truncated by the shear; similar dykes in the shear zone are folded with their host rocks.*

is evidence to support a localized contact thermal overprint from the NPS. It is thus not certain if granulite-facies assemblages in gneisses near the plutons reflect regional or contact effects, or a combination of both.

The dykes in both the Nukasusutok and South Aulatsivik areas record evidence for both granulite- and amphibolite-facies metamorphism, with mineral assemblages that mirror those present in the Dog Island–Ford Harbour area (Ryan, 1991a). The problem of the origin of the granular orthopyroxene-bearing dykes (i.e., are they metamorphic or igneous rocks ?) was outlined last year, and is no more easily resolved based on the present work. It is quite apparent that such dykes are both massive and foliated, and some of them follow the shear zones mentioned above. Assemblages may in some cases be related to proximity to the NPS, but this correlation has not been firmly established.

The gneisses and deformed plutonic rocks south of Nain Bay are cut by many east–west, largely sinistral, faults that displace contacts tens to hundreds of metres. Northeast- and northwest-trending faults in some cases may be splays of the east–west structures, but in other cases seem to be an unrelated set. The gneisses to the west of the deformed norite–anorthosite generally dip steeply beneath the latter (Figure 4). Along the northern junction, however, the contact is shallow, and the layered melanorite unit seems to define an overturned fold structure plunging gently to the southwest. An interesting link between most of the rocks in this area is a regionally developed shallow- to moderately-south plunging mineral-streaking lineation. The attitude of this lineation is unlike that which occurs in gneisses of the Churchill Province to the west, the latter being generally subhorizontal. The fact that the streaking lineation is recognizable within gneisses, anorthositic rocks and the granitoid rocks is suggestive of a common origin.

The region south of Nain Bay was interpreted by Ryan (1990), to contain the Nain–Churchill (Rae) boundary. This was based on the geological mapping of E.P. Wheeler, whose manuscript maps showed a distinct change in rock types within the gneissic terrane of this area (see discussion in Ryan, 1991a). There now seems to be no firm basis for this interpretation. If a tectonic assignment has to be made, the writer now favors inclusion of all rocks here within the Churchill (Rae) Province, since no obvious structural break exists between any of the three gneissic units in this area.

MINERALIZATION

The quartzofeldspathic gneisses appear to have little mineralization. A narrow rusty zone in which coarse-grained pyrite and/or pyrrhotite occurs, was discovered in a mafic granulite by S. Swinden, of the Mineral Deposits Section, during a brief visit to the field area in August. The mineralization occurs near a ductile shear zone on one of the northwestern islands of the Akulaitut group.

Dense, rusty, ilmenite–sulphide-bearing stratiform zones occur within well-layered rocks on a small island just east of Nukasusutok (Figure 2; Plate 3b). These are the southernmost exposures of layered, deformed, and recrystallized norite that, according to Davies' (1974) map, extends northward across Nukasusutok Island proximal to the eastern margin of the massive anorthositic rocks that underlie most of the island. More mineralized rocks of this type are thus probably present northward along strike.

The largest mineralized zone found during the 1991 season occurs within a 5- to 10-m-wide pegmatoidal north-trending norite dyke that intrudes recrystallized, equigranular anorthosite and banded norite and melanorite on the western margin of the gneissic anorthosite–norite unit south of Nain Bay (Figure 4). Crystal faces of pyrrhotite on the order of several centimetres in size are visible in the pegmatite from the least indurated parts of the gossan zone. In some cases, the sulphide appears to be a surface encrustation or replacement of coarse orthopyroxene, whose morphology in the sulphide-free pegmatoidal rocks is identical to that of the coarse sulphide. Chalcopyrite occurs interstitial to the pyrrhotite–pyrite in some parts of the showing. Although this is the most spectacular of the rusty zones in the anorthositic rocks of this area, it is not the only occurrence of this type. Several other rusty ilmenite-bearing leuconorite dykes intrude foliated norite and sugary recrystallized anorthosite to the east, and locally form stockworks that exceed the amount of older rock in outcrop. In addition, Grimley (1955) has reported similar mineralization in 'small shear zones in anorthosite country rock' from the south shore of Nain Bay, which may be of the same age and origin as that described above.

SUMMARY AND DISCUSSION

The 1991 mapping has shown that the general subdivisions of the gneiss complex established last year for the Dog Island–Paul Island area can be extended southward to the Nukasusutok Island area. Northward, however, the gneisses are not easily subdivided, in part because the granulite-facies metamorphism makes subtle differences in the quartzofeldspathic rocks difficult to detect. As in the Dog Island area, as well, supracrustal rocks are not areally extensive at the 1:50 000 map scale. The paragneisses are predominantly quartzite (or perhaps recrystallized chert) akin to those around Ford Harbour. The mafic granulites and amphibolites distal from the plutonic rocks are typically black, hornblende-rich blotchy-textured rocks, whereas those adjacent to plutons (such as the Jonathan Island Intrusion) are equigranular pyroxene-rich rocks. The latter may be related to the adjacent plutonic rocks in some way, or their pyroxene-rich (and commonly olivine-bearing) character may be a function of contact metamorphism. The Jonathan Island granulite with its layers and sheets of purplish-grey granular norite is a typical example of these mafic gneisses, and can be used to illustrate the problems they pose. This unit was considered by Berg and Breigel (1983) to be of 'sedimentary origin', but they were somewhat perplexed by the igneous composition of the grey rock, especially where it occurred as a matrix to a fragmented white anorthosite. They concluded the brecciated anorthosite was probably a pyroclastic rock, a conclusion that is not favoured by the present writer. It seems more likely that the mafic granulite and white anorthosite are older igneous rocks that have been brecciated by intrusion of the grey-purple norite, and perhaps the whole assemblage is related to the Jonathan Island intrusion. White anorthosite layers are also present in the layered mafic rocks on Carey Island, suggesting continuity of this 'granulite' unit intermittently around the whole of the Jonathan Island intrusion (cf. Ryan, 1990).

Metamorphosed and deformed basic dykes occur in both the northern and southern areas of gneisses mapped during the 1991 season. As in the Dog Island area, there are several generations of such dykes, and they display morphological characteristics that indicate intrusion under brittle and ductile conditions of their hosts. Both granular and foliated dykes are present, exhibiting mineralogical characteristics of amphibolite- and granulite-facies metamorphism. Subophitic and porphyritic textures are recognizable in some dykes, but even these dykes have a metamorphic overprint. The attitude of the dykes in the Nukasusutok Island–Humbys Island area changes from subvertical in the west to subhorizontal in the east, possibly indicative of increasing distance from source eastward. The lamprophyric dykes are an intriguing suite that seem to be unique to the Nain area. Their absolute age has not yet been determined, but if related to NPS magmatism, they represent a newly recognized member of this anorogenic association.

Deformed anorthositic and granitic rocks that were discovered in 1990 south of Nain Bay (Ryan, 1991a) were investigated in more detail during the 1991 field season and found to be important components of the Nain Plutonic Suite in this area. These rocks present some of the most intriguing problems within the Nain region. Deformed and recrystallized anorthositic and noritic rocks form at least two geographically separate units south of Nain Bay, and one of these is intruded by massive pegmatoidal leuconorite. In all likelihood, more such zones of deformed anorthositic rocks occur within the NPS, as is suggested by the examination of a 'granulite of uncertain origin' on the south shore of Kingurirtik Lake. Ferrodioritic to monzonitic rocks in the Nain Bay area form dyke-like bodies intruded into already deformed anorthosite, yet are themselves strongly deformed. Unlike the non-foliated granitoid rocks of the NPS, which form somewhat bulbous and irregular intrusions plainly discordant to regional trends (see Ryan, 1990) the dyke-like bodies south of Nain Bay are concordant with the regional trend of the gneisses, clearly indicating a structural control on emplacement. It was the elongate shape of these granitoids on E.P. Wheeler's manuscript maps that in part prompted the writer to investigate this area.

The results of the 1991 mapping outlined above reinforce the conclusion from the 1990 work that the geological history of the Nain area is probably more complex than hitherto appreciated. Facts that point to this were discussed by Ryan (1991a)—specifically, the several generations of dykes and their metamorphic state, and the definition of deformed and recrystallized rocks within the NPS. Each of these will be briefly discussed in light of the results of the 1991 work.

The small database of precise dating from rocks in the Nain region has proven a hinderance to evolutionary modelling, so any proposal for its geological history is open to different interpretations as will be evident below. This is especially so with the basic dykes, for which no firm absolute ages are yet available. As noted earlier (Ryan, 1991a), certain dykes in the wispy granitoid gneiss of the Dog Island area appear to be syn-plutonic with the host. Preliminary zircon U–Pb data by J. Connelly (personal communication, 1991) of Memorial University of Newfoundland suggest that a late Archean age (ca. 2.54 Ga) is probable for both the gneiss and a dyke at one such outcrop, indicating a hitherto unknown period of basic dyke intrusion in the Nain Province. It is possible that some of the other dykes are also Archean. However, by comparison with the Hopedale and Saglek blocks, others are quite probably Early Proterozoic. This raises the (unlikely) possibility of Hudsonian deformation having encroached on this area as noted in a discussion of the 1990 mapping data (Ryan, 1991a, p. 248). However, if the assumption is made that at least some of the dykes with plagioclase phenocrysts and noritic fragments must be related to NPS magmatism, then foliated Middle Proterozoic dykes are present as well, and the tectonism of this age must be an effect of the emplacement of the plutons. It is unlikely that the sigmoidal foliations in such dykes result from magmatic flow, but instead attest deformation imposed upon the dykes as a result of post-emplacement stress. Foliations of this type are a result of lateral slip of the dyke walls either just prior to final and complete crystallization or are a result of post-consolidation reactivation of the dyke as a plane of shear (cf. Berger, 1971; Davidson and Park, 1978).

The deformed and recrystallized igneous rocks that form units distinct from adjacent gneisses on the islands also pose more questions and lead to alternative scenarios in reconstructing the geological history of the area. If, for instance, the melanocratic norite–monzonite association containing leuconorite inclusions on Satok Island is Late Archean instead of Middle Proterozoic in age, then the deformation that has affected it could be Late Archean or Hudsonian. This of course implies the existence of norites and leuconorites of Archean age that are mesoscopically indistinguishable from the deformed rocks of the areally more-extensive NPS. If it is of the same age as the Bridges layered intrusion (ca. 1650 Ma, L. Ashwal, personal communication, 1991), the deformation must be Middle Proterozoic but could be of Labradorian (1650–1620 Ma) rather than Elsonian (1320–1200 Ma) age. Likewise, if these igneous rocks are an integral part of the 1320–1290 Ma magmatism of the NPS, their foliation must have been impressed during Middle Proterozoic deformation. The same uncertainty of place must be used when discussing the deformed lamprophyric and kimberlitic dykes that intrude the monzonite and also intrude the gneisses on the nearby islands. These could similarly be Late Archean or Middle Proterozoic in age. By inference, the shear zones that truncate such dykes on Nuasurnak Island could be of Late Archean, Early Proterozoic, Labradorian, or Elsonian age! All these examples illustrate the problems encountered when tectonic models for the Nain area are contemplated.

The layered noritic gneiss of eastern Nukasusutok Island is another example of a rock whose relative position in the stratigraphy of the Nain area is unclear. E.P. Wheeler's sense of uncertainty about the origin of this rock is understandable. This 'granulite' differs from the hornblende-rich mafic gneisses of the main gneiss complex by (1) being noritic in composition, (2) having vestiges of primary igneous textures, (3) having a primary compositional layering, albeit enhanced by deformation, (4) lacking the granitic network and orthopyroxene-bearing 'sweats' that occur in the mafic rocks to the east, and (5) having abundant ilmenite, a mineral that is uncommon in the mafic granulites to the east. This rock thus has more in common compositionally, mineralogically, and texturally with the Middle Proterozoic plutonic rocks of the area than with the gneisses and, in fact, is texturally comparable to some of the deformed and recrystallized noritic rocks that occur south of Nain Bay. Its deformation, including ductile shear zones and isoclinal folds, sets it apart from the directly adjacent undeformed plutonic rocks, but it seems difficult to escape the conclusion that this rock was derived from a parent akin to or directly related to the NPS. A probable pre-NPS equivalent could be the Bridges layered intrusion, a locally deformed olivine-bearing gabbroic to noritic body near Nain that predates the emplacement of the adjacent NPS anorthosites (Planansky, 1971) and may be as old as 1650 Ma (L. Ashwal, personal communication, 1991).

Deformed and recrystallized igneous rocks that are presently considered to belong to the NPS are also problematical. Their recognition in 1990 debunked the widely-held belief that the NPS is a pristine anorogenic terrane that exhibits no significant deformation. The recognition of penetrative tectonic fabrics and recrystallization textures within the NPS, therefore, prompted Ryan (1991a, c) to examine two current alternatives to account for their presence. The two potential mechanisms for the tectonic overprint are (i) diapiric emplacement of semi-solid cumulates or (ii) penetrative deformation following solidification, totally divorced from the emplacement itself. Consideration of these models assumes the deformed anorthositic rocks are not older than the NPS. An Archean or Early Proterozoic age however is not entirely ruled out for some at this stage of investigation, since similar rocks appear to be part of the gneiss complex farther south (Ryan and Lee, 1986).

Deformed noritic and anorthositic rocks of the type found recently in the NPS were recognized decades ago in anorthositic rocks in Norway. There they appear to be confined to the margins of individual intrusions and their foliation attitude implies they form an external sheath to domal bodies (cf. Michot, 1969). They have been interpreted by J.C. Duchesne and colleagues (cf. Duchesne and Michot, 1987; Duchesne and Maquil, 1987; Maquil and Duchesne, 1984; J.C. Duchesne, personal communications, 1991, 1992) to be a product of marginal deformation of the plutons by frictional drag during their emplacement as crystal mushes. Inclusions of foliated anorthosite–norite within interior undeformed rocks are interpreted to be related to late-stage puncturing of the gneissic carapace by the rising magmatic core and root of the diapir, which is now expressed as the central part of the domes. Using the Duchesne model, the gneissic rocks south of Nain Bay (Figure 4) could be interpreted as the deformed margin of the massive leuconorite pluton to the east, the latter being the core of the diapir that ascended and pierced the enveloping gneissic sheath during the final expansion and flattening of the pluton. This model has at least one feature that can make it applicable to the Nain Bay area and elsewhere where similar rocks have been observed (cf. Ryan, 1991a), namely, the position of the deformed plutonic rocks relative to the main plutonic masses. Most of the strongly deformed anorthosite and norite seen to date by the writer occurs proximal to gneisses, suggesting a marginal locus for the foliated rocks. However, this observation may be biased by the focus of the project, namely the gneisses themselves; consequently, it is only near these gneisses that the plutonic rocks have been examined systematically. It is quite obvious from the Kingurutik Lake occurrence of gneissic norite that such zones also occur in the interior of the NPS.

There are a few features of the Nain Bay anorthosites that suggest a simple domal diapir model is not applicable. For instance, the country-rock gneisses locally dip gently beneath the gneissic norite–anorthosite rather than away from it. Thus, if the intrusion is diapiric, the foliated plutonic rocks of this area must represent a section through a part of the deformed carapace that has spread laterally above the gneisses in a mushroom-like pattern (cf. Jackson and Talbot, 1989). This implies nappe formation or oblique ballooning of the kind suggested, for example, by Martignole and Schrijver (1970) to have been operative during diapiric emplacement of the Morin anorthosite, Quebec, and by Courrioux (1987) for diapiric emplacement of the Criffel granodiorite–granite pluton in the Southern Uplands of Scotland. Lateral migration of anorthosite away from a semi-solid diapir as a nappe-like mass at Morin has, however, been rejected by Emslie (1975) who argued that the deformed rocks of the Morin intrusion are a consequence of post-solidification regional tectonism.

The presence of dyke-like masses of foliated diorite–monzonite in the western part of the Nain Bay map area, as noted above, is a pattern uncharacteristic of most of the other ferrodioritic and granitoid rocks of the NPS. The Martignole and Schrijver (1972) model would assign these as magmas that co-existed with a largely crystalline anorthositic diapir, essentially forming a 'liquid cap' above the diapir, and would be interpreted as keels of these sheets now infolded or squeezed into the roof of the diapir. Alternatively, they could be dykes injected downward into fractures in the anorthosite (Emslie, 1975, p. 33). Several problems arise with these interpretations. Although these intrusions do occur within anorthosite, others occur totally within the country-rock gneisses, and therefore, are not part of a cap singularly associated with the anorthosite of this area. Under the above models, however, the dioritic rocks could perhaps be accommodated as downward tapering wedges expulsed obliquely from the diapirs, with a morphology akin to curved flange intrusions (cf. Walker, 1975).

The sheeted contacts noted earlier that mark the junction between the diorite and anorthosite could represent regions where forceful downward injection of magma resident above the diapir has taken place, as suggested by the models referred to above. However, these contacts are here interpreted as zones in which vertical *upward* intrusion of the younger rocks was abnormally intense–the writer knows of no contacts between anorthosite and the large bodies of granitic rocks within the NPS that are of this type. Perhaps this signifies a greater depth of erosion than for the other granitoids of the area, and may in fact be a reflection of features that characterize the deeper contacts between all the anorthositic and granitic rocks within the NPS. These sheeted zones and dyke-like bodies could thus represent the migration routes for ascending magma that formed the more irregular granitic plutons that represent higher levels of the igneous terrane (cf. Morse, 1969b). Apparently, marginal zones between anorthositic and gneissic rocks to the west and north are marked by similar dyke networks (Wheeler, 1969).

Another problem that arises in the consideration of the diapiric model is that it requires the diapir to comprise a lubricating interstitial liquid (of noritic composition?) between crystals of plagioclase and orthopyroxene. As noted last year, evidence for existence of this liquor seems to be lacking, and the favoured interpretation is that all the deformation seen in the NPS is imposed upon already crystalline masses (see also Emslie, 1975, p. 31). This is obvious from observations that subophitic intergrowths of plagioclase and orthopyroxene are deformed, implying solid-state deformation. Diapiric rise of a solid anorthosite–norite body seems to be a mechanism for which little information is available, but Emslie (op. cit.) notes that significant buoyant uprise of igneous bodies is inhibited once advanced solidification has occurred. Subvertical flow alignment of the plagioclase and orthopyroxene is also unknown; such preferred orientations should be present in mobile, diapiric crystal mushes.

A dichotomy of the diapiric model of anorthosite emplacement within NPS is that only certain plutons show the foliated zones (usually interpreted as margins) attributed to synemplacement deformation of semi-consolidated masses. If anorthositic terranes are constructed via coalesence of nearly solid crystal mush diapirs, rather than by intrusions of plagioclase saturated liquids (Wiebe, 1990) or liquid-enriched crystal-laden magma, then why do only certain plutons display the foliated margins? It would seem logical to assume, using solely the crystal mush diapir model, that all anorthositic plutons should have gneissose zones separating them from adjacent country rock and from older components of the NPS itself. This is plainly not the case, as is clearly demonstrated by the many igneous (intrusive) contacts that have been described from within the NPS. Similarly, in the Rogaland massifs of Norway, even though the Egersund–Ogna has a foliate margin and is interpreted as a diapir, the younger Helleren massif does not (Duchesne and Michot, 1987, p. 49). Wiebe (1991) acknowledged such differences between plutons and proposed different mechanisms of intrusion for plutons of the NPS on Paul Island near Nain (Figure 1) that may account for morphological variation between plutons in such terranes. He concluded that the diapiric plutons–i.e., those to which he assigned a foliated carapace–are the early members. He observed that high-alumina orthopyroxene megacrysts appear to be restricted to this type, and he reasoned that such intrusions represent crystal mushes that rose 'from depth', prior to emplacement of the undeformed massive plutons, which are products of magma crystallization in upper crustal chambers. If this equating of the presence of orthopyroxene megacrysts and the degree of deformation with the depth of crystallization is correct, then it implies that potentially a major part of the NPS has been intruded diapirically from a deeper level. This conclusion is based on observations of the distribution of coarse orthopyroxene and deformed rocks, in conjunction with the re-interpretation of E.P. Wheeler's 'granulite of uncertain origin' (equated now with foliated norite) on the margins of the NPS (cf. Ryan, 1991a). The distribution of the 'granulite', based on its distribution on Wheeler's maps, suggests that the whole anorthositic core of the NPS from Anaktalik Bay to Okakh Bay (cf. Ryan, 1990) has been emplaced diapirically. If this is so, it represents an enormous volume of plagioclase that rose through the crust after the plutons had magmatically amalgamated at some greater depth.

The shallow to moderate plunge of the lineation within a generally steeply-dipping foliation plane in gneisses, noritic rocks and granitoid rocks in the area examined south of Nain Bay (Figure 3) does not appear compatible with subvertical transportation as implied by diapiric tectonics, but is more strongly suggestive of oblique-slip within both the gneisses and the igneous rocks. If diapiric tectonics are responsible, then movement has been strike-slip as well as dip-slip, and the pluton(s) responsible for this type of tectonic transport must have been subjected to lateral ballooning of an already solid margin as noted earlier. A problem with this model, however, is that the elongate diorite–monzonite plutons contain inclusions of foliated anorthosite-norite, and many smaller dykes within the western anorthositic mass south of Nain Bay crosscut a foliation in their host, so solid-state deformation of the basic rocks occurred before emplacement of the granitoids, after which further deformation affected both. Under the Emslie (1975) and Martignole and Schrijver (1972) models the diorite–monzonite intrusions represent infiltrations of the liquid cap into the solid, but fractured, carapace. It is difficult to percieve how actively ballooning anorthositic diapirs under ductile compressive stress from internal magma pressure could permit such large dykes of granitic material to penetrate them. Another problem with the diapir model is that there seems to be a lack of any subvertical linear fabric in the 'piercing plutons', which could potentially have been the rising diapir core (cf. Jackson and Talbot, 1989); as noted above, some primary alignment of crystals reflecting flow should be obvious in these rocks. In addition, plutons that appear to be candidates for the ascended diapir core (for instance, the pegmatoidal leuconorite south of Nain Bay; Figure 4) still display a megascopic-scale subhorizontal layering, not the steep chaotic structures of the type modelled in experiments of diapirism of stratified (i.e., already cumulate-layered, semisolid) bodies.

Ryan (1991a, c) has suggested previously that another mechanism may be responsible for the formation of the deformed rocks within the NPS, namely that they represent zones of shear that reflect episodic ductile faulting during the emplacement period of the NPS. Under this model the anorthosite–country-rock contacts and the elongate granitoid plutons (and smaller dykes of the same) become preferred zones of shear during imposed strain because of compositional contrast. The steep attitude of the foliation in all units indicates the shears are subvertical, except immediately south of Nain Bay where shallow dips and an apparently overturned fold suggest northward translation of the gneissic anorthosite over the quartzofeldspathic granulites. Regardless of layering attitudes in this region, the lineations are consistently south-plunging indicating a generally north-south movement direction. The fold may therefore be a function of re-activation of an original gneiss-pluton contact that was at a high angle to the shear direction. It is conceded that the proposed model has weaknesses, and is based on limited examination of the plutonic rock.

As noted earlier, deformed (foliated and lineated) ferrodioritic–monzonitic rocks contain rafts of earlier-deformed and recrystallized norite. The implication of this relationship is that the internal and marginal shear zones of the anorthositic rocks become zones of intrusion for the granitoids. These dyke-like bodies themselves then become the locus of movement in the same manner as the basic dykes in the country-rock gneisses as noted above. The latter feature is apparent in many of the smaller dykes of monzonite–ferrodiorite within the anorthositic rocks.

It is perhaps possible, as a first correlation, to link the development of ductile shear zones in the gneisses of the archipelago with the development of foliations and lineations in the igneous rocks south of Nain Bay. As noted by Ryan (1991a), it is not unreasonable to advocate episodic movement along subvertical fault zones during emplacement of the NPS. The presence of high-temperature ductile shears in the gneisses may be evidence of such fault zones. Their sites of generation mirror the high-temperature environment of the foliated and recrystallized rocks within the NPS; in the latter, orthopyroxene, clinopyroxene, and olivine are stable minerals. The gneisses and plutonic rocks thus record high strain at high temperatures along oblique-slip ductile translation zones. The high-temperature assemblages in these deformed rocks may reflect the ambient conditions of the crust due to the magma pooling underneath. If the proposal that shear zones in the gneisses and the deformed and recrystallized plutonic rocks of the NPS are a consequence of zonal deformation during the Mid Proterozoic is valid, then these features are a result of the tectonic regime that prevailed at the time of NPS emplacement. These indicate that subvertical oblique movements were dominant over subhorizontal crustal extension. It has already been suggested (Ryan, 1991b) that pluton emplacement and the location of shear zones within the NPS may be governed by ancestral structures originally developed in the underlying crust during collision of Nain and Churchill (Rae) provinces ca. 1800 Ma.

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GEOLOGY OF THE ALEXIS RIVER MAP REGION, GRENVILLE PROVINCE, SOUTHEASTERN LABRADOR

T. van Nostrand, D. Dunphy and D. Eddy
Labrador Mapping Section

ABSTRACT

The Alexis River map region, located in the Grenville Province, southeastern Labrador, is divided into three fault-bound crustal segments. These are the White Bear Arm complex, the Gilbert River belt and the Mealy Mountains terrane.

The White Bear Arm complex comprises gabbro and gabbronorite (consisting of original igneous textures and granulite-facies rocks having metamorphic textures) and enclaves of metasedimentary gneiss and granite. The Gilbert River belt consists of a series of northwest-trending belts of orthogneiss, metasedimentary gneiss, foliated K-feldspar megacrystic granite to granodiorite and non-megacrystic granite to granodiorite, and an extension of the Alexis River anorthosite. The Mealy Mountains terrane is underlain by variably foliated K-feldspar megacrystic granodiorite to granite, diorite, granodiorite and amphibolite orthogneiss, pelitic and psammitic metasedimentary gneiss and a large composite intrusion comprising variably recrystallized and deformed syenite to quartz monzonite (informally named the Upper Paradise River pluton).

Structure in the region is dominated by a northwest regional trend in the Gilbert River belt and a variable east to northwest trend in the Mealy Mountains terrane. Metamorphism attained upper amphibolite facies and there is evidence that granulite-facies conditions were locally reached.

INTRODUCTION

During the 1991 field season, systematic 1:100 000-scale mapping was carried out in the Alexis River region, within the Grenville Province, southeastern Labrador. The region comprises NTS 1:50 000-sheets 13A/11, 12, 13, and 14, and encompasses an area of approximately 3700 km^2 (Figure 1). Previous geological mapping within the region is included in a 1:500 000-scale geological map of Eade (1962). Based on isolated spot landings, he describes the area as being underlain predominantly by undifferentiated granitoid rocks, orthogneiss, and monzonitic to granodioritic rocks. Much of the western portion of the region has remained unmapped until the present study. Complete aeromagnetic map coverage at 1:250 000 and 1:50 000 scale is available for the region (Geological Survey of Canada, 1974a, b). A 1:250 000-scale, colour-shaded relief aeromagnetic map is also available for the region (Kilfoil, 1991). The adjacent Paradise River and Port Hope Simpson map regions to the north and east of the study area were mapped at 1:100 000 scale by Gower *et al.* (1985, 1987).

The western half of the region occupies a large plateau underlain by extensive fluvioglacial deposits, including eskers, dunes and sand plains. The central portion consists of large string bogs, muskeg and heavily wooded areas. Bedrock exposure is limited to elevated areas and along river valleys. The eastern portion, through which the Alexis River (and after which the map area is named) flows, is well exposed, particularly in the southeast corner, where extensive burned-over areas occur.

REGIONAL SETTING

The Alexis River map region is located within the Grenville Province, southeastern Labrador, and straddles three major lithostructural terranes (the Hawke River, Lake Melville and Mealy Mountains terranes, Figure 1; Gower *et al.*, 1987). The terranes are distinguished by different metamorphic, structural and radiometric characteristics. The distribution of rock units in the Alexis River map region is shown in Figure 2. The area has been divided into three major zones, following Gower *et al.* (1987). These include from north to south; the White Bear Arm complex (part of the Hawke River terrane), the Gilbert River belt (included in the Lake Melville Terrane), and the Mealy Mountains terrane. These zones are continuations of units mapped in the Paradise River and Port Hope Simpson map regions to the north and east, respectively, (Gower *et al.*, 1985, 1987). Geochronological data are not available within the area, but

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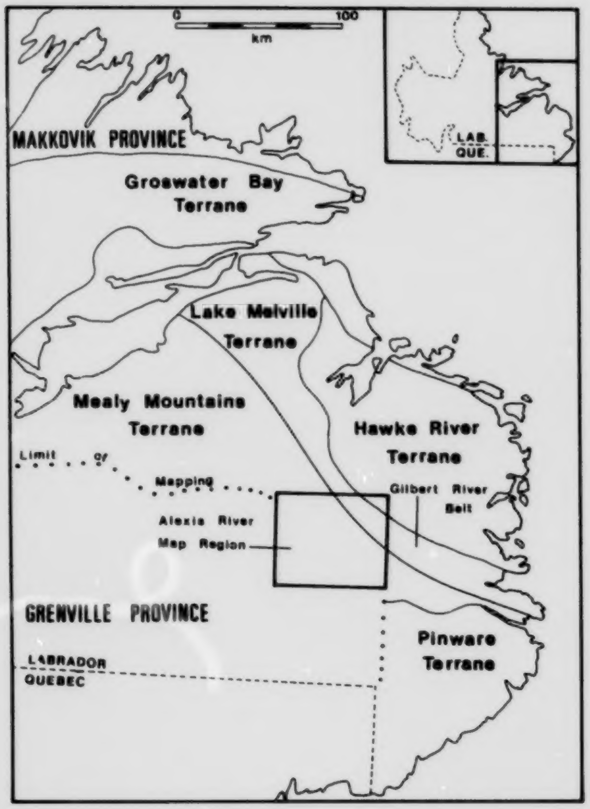


Figure 1. *Regional subdivision of the Grenville Province in eastern Labrador (after Gower* et al., *1988a), and location of the Alexis River map region.*

extrapolating from dated units in the above two regions suggests that the most of the rocks in the study area have ages between 1680 and 1630 Ma.

Figure 2 is a simplified version of a 1:100 000-scale map compiled from approximately 900 data stations. All samples collected were slabbed and stained (specifically for K-feldspar). Unit designators on Figure 2 follow the legend of Gower *et al.* (1988b, c). Age relations are not necessarily implied by the order of the number groups. For example, Unit 1 (orthogneiss) rocks are in some cases only strongly deformed equivalents of Unit 4 (foliated granitoid rocks). Nor are age relations implied by the order in which units are presented within each number group. Units 3 and 5 are omitted from use in the Alexis River map region, as these units are reserved for rocks of the Earl Island domain (Unit 3) farther north and presumed syn-Grenvillian granitoid rocks (Unit 5) to the east in the St. Lewis River map region.

WHITE BEAR ARM COMPLEX

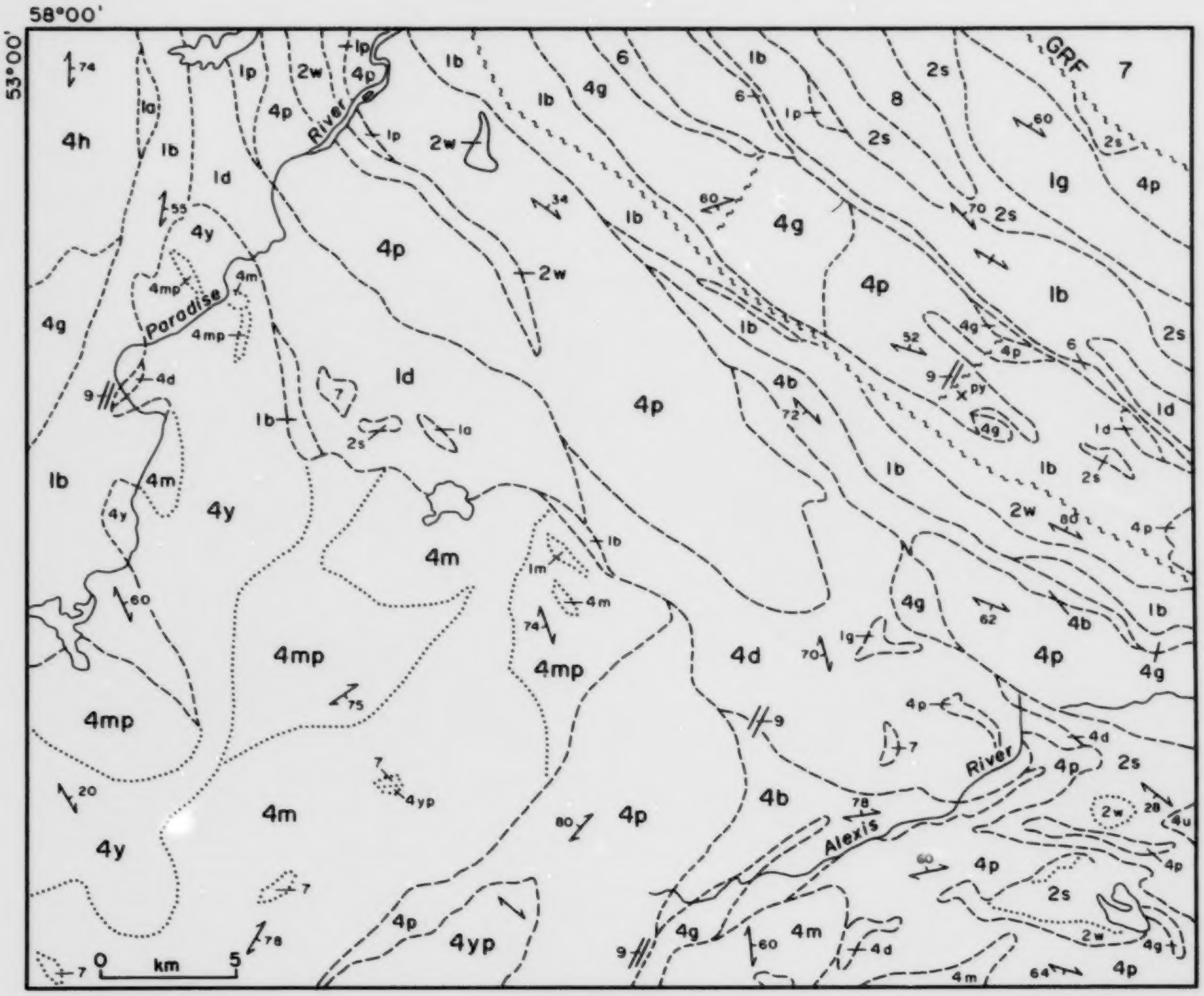
The extreme northeast corner of the map area is underlain by a small segment of the White Bear Arm complex (Unit 7), a large layered mafic intrusion mapped from Sandwich Bay to the southeast coast of Labrador, a distance of 150 km (Gower *et al.*, 1987). Within the area, the White Bear Arm complex comprises gabbronorite having enclaves of K-feldspar megacrystic granite, metasedimentary gneiss and mylonitic granite. The gabbronorite consists of a major and a minor component. The major rock type is a homogeneous, grey-weathering, medium-grained massive rock containing a mauve-coloured plagioclase, light-brown orthopyroxene, light-green clinopyroxene, olivine, hornblende and an opaque mineral. Original (igneous) minerals and textures are commonly preserved, although recrystallized and hydrated metamorphic equivalents occur. Pyroxenes (ortho and clino) are partially replaced by amphibole and, locally, coronitic textures, consisting of olivine cores having orthopyroxene and amphibole rims, are developed. The minor component of the White Bear Arm complex is a grey- to brown-weathering, fine- to medium-grained, equigranular, recrystallized granulite-facies norite. Minerals include a white to grey plagioclase, light- to dark-brown orthopyroxene, pale-green clinopyroxene and an opaque mineral. These rocks occur as small, isolated outcrops within the gabbronorite. Gower *et al.* (1987) interpret similar rocks in the Port Hope Simpson region as feeder dykes for the later gabbronorite.

Within the White Bear Arm complex gabbro, three types of enclaves are present. The first type is a pink- to red-weathering, fine- to medium-grained granite, having a well-developed mylonitic C−S fabric (Plate 1). Minerals present include recrystallized quartz, plagioclase, K-feldspar, garnet, biotite and an opaque mineral. At one locality, a two-pyroxene norite dyke intrudes the granite. The second type of enclave is a white-weathering, strongly recrystallized and foliated, granitic-textured rock containing quartz, plagioclase, K-feldspar, an opaque mineral and garnet. These rocks are similar in appearance to anhydrous metasedimentary gneisses found in contact with the Sand Hill Big Pond gabbronorite (Gower *et al.*, 1986). The third type of enclave is a medium-grained, moderately foliated, K-feldspar megacrystic granite, containing partially recrystallized K-feldspar megacrysts, and a groundmass consisting of quartz, plagioclase, K-feldspar, biotite, hornblende, and an opaque mineral.

GILBERT RIVER BELT

The southern margin of the White Bear Arm complex is marked by an extension of the Gilbert River fault (described by Gower *et al.*, 1987). The fault is a major northwest-trending structure, which marks the boundary of the Hawke River terrane to the north and the Lake Melville terrane to the south. Although not exposed in the Alexis River region, the fault is indicated by a strong, low-aeromagnetic signature and airphoto lineament.

The Gilbert River belt within the region is a 30-km-wide belt comprising foliated to gneissic granite to granodiorite, sillimanite-bearing metasedimentary gneisses, K-feldspar megacrystic granite and a narrow segment of the Alexis River anorthosite (which includes associated dioritic gneiss). Gower *et al.* (1987) proposed the name the Gilbert River shear belt for the extension of these rocks in the Port Hope Simpson



LEGEND

LATE PRECAMBRIAN–EARLY PALEOZOIC
9 *Gabbro, diabase*

MIDDLE TO LATE PROTEROZOIC
8 *Pegmatite*

MAFIC INTRUSIVES
7 *Gabbro, gabbronorite*
6 Alexis River Anorthosite

GRANITOIDS
Upper Paradise River Monzonite
4mp *K-feldspar megacrystic monzonite to quartz monzonite*
4yp *K-feldspar megacrystic alkali feldspar syenite*
4y *Syenite to quartz syenite*
4m *Monzonite to quartz monzonite*

GRF–Gilbert River fault

MIDDLE TO LATE PROTEROZOIC
4p *K-feldspar megacrystic granite to granodiorite*
4g *Granite, alkali feldspar granite*
4u *Muscovite-rich granite*
4b *Granodiorite*
4h *Hornblende syenite*
4d *Diorite to quartz diorite*

PARAGNEISS
2s *Sillimanite-bearing pelitic metasedimentary gneiss*
2w *Psammitic metasedimentary gneiss*

ORTHOGNEISS
1a *Amphibolite gneiss*
1g *Granite gneiss*
1b *Granodiorite gneiss*
1d *Diorite gneiss*
1m *Monzonite gneiss*
1p *K-feldspar megacrystic granite gneiss*

SYMBOLS

--- ······ *Geological contact (approximate, assumed)*
~~~~~ *Undifferentiated faults (mostly strike-slip)*
↔ *Gneissosity, foliation*
×py *Pyrite occurrence*

Figure 2. *Geology of the Alexis River map region.*
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CENTIMETRE

Plate 1. *Mylonitic granite showing C–S fabric suggesting dextral (west over east movement); enclave in gabbro of White Bear Arm complex. S–foliations; C–shear band.*

map region to the east. However, subsequent detailed mapping of this belt of rocks on the coast of Labrador by Hanmer and Scott (1990), suggests that although the belt is a zone of strong deformation, the proportion of intensely strained rocks within the coastal section is small, and preferred the name Gilbert River belt to describe the zone. In the Alexis River map region, most of the units within the belt are strongly deformed; however, in the northwest part of the belt, deformation is generally only moderate, and zones of intense deformation are very localized. Thus, the name Gilbert River belt will be used throughout this report.

Most rocks within the Gilbert River belt are extensions of units mapped in the Paradise River and Port Hope Simpson map regions (Gower *et al.*, 1985, 1987). All rocks within the zone have a variable northwest-trending fabric, excluding some minor dykes that are probably late-stage intrusions.

The oldest rock unit is a pelitic, sillimanite-bearing metasedimentary gneiss (Unit 2s). The paragneiss occurs as a Y-shaped wedge continuous across the northwest corner of the region and as enclaves (m to km scale) within other units of the Gilbert River belt. Three types of metasediments are present. The first type is a pink- to brown- to cream-weathering, schlieric to layered migmatite, exhibiting complex and variable fold styles. The fabric is defined by alternating quartz, feldspar-rich leucosome and a biotite–sillimanite–garnet–opaque mineral-rich melanosome. Retrograde muscovite is locally present and typically postdates the time of fabric formation. Garnet and sillimanite are commonly present in the leucosome, probably as melt products. Within the wedge-shaped body of pelitic gneiss immediately south of the Gilbert River fault, veins of K-feldspar megacrystic granite contain small enclaves and schlieren of sillimanite-bearing gneiss. These commonly contain only the restitic phase, suggesting a high degree of assimilation of the pelitic gneiss by the intruding megacrystic granite. The second type is a white- to rusty-weathering homogeneous, diatexitic rock, which occurs as gradations from the well-layered paragneiss. Garnet, biotite, sillimanite and opaque mineral assemblages are present, typically as thin, discontinuous restitic seams. The third rock type is a well-banded, medium- to coarse-grained, typically garnetiferous psammite that is also present as small-scale intercalations within the pelitic gneiss. This rock type is subordinate amphibolitic gneiss associated with the metasediments, and minor K-feldspar megacrystic granite are present as kilometre-scale enclaves.

A major rock type within the zone is a pink- to red-weathering, granodiorite gneiss. The gneiss varies from fine grained and well banded to medium grained and diffusely banded. Banding is defined by alternating quartz + K-feldspar + plagioclase layers, and biotite + hornblende + garnet + opaque phase-rich segregations. Small, concordant granite and alkali-feldspar granite pegmatite veins are common, and in many areas enhance the gneissic appearance (Plate 2). In some areas, gradations from well-banded gneiss to foliated granodiorite on outcrop and kilometre-scale suggest inhomogeneous strain within the Gilbert River belt. Several enclave types are present within the gneiss, including sillimanite-bearing pelitic gneiss, K-feldspar megacrystic granite, weakly foliated to massive granite, and dioritic gneiss.

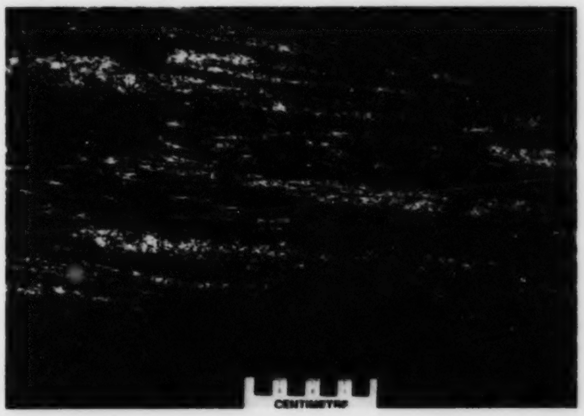

Plate 2. *Banding in granodiorite gneiss enhanced by concordant granite veins (Gilbert River belt).*

Granite gneiss (Unit 1g) occurs as a wedge-shaped body in the northeast corner of the region. These rocks are typically pink weathering, fine to medium grained, recrystallized, and well banded. Banding is defined by alternating quartz, plagioclase, K-feldspar-rich layers and biotite, hornblende, garnet and opaque mineral-rich layers (Plate 3). Gradations into a more granodioritic composition occur on outcrop scale, although not depicted on the map. In some areas, the gneissosity is enhanced by intruding, concordant, locally complexly folded granitic veins. Discordant granite veins and coarse pegmatites are a minor, but common, intrusive phase in this unit. Enclaves within the gneiss include amphibolitic gneiss, psammitic metasediments, foliated granodiorite and K-feldspar megacrystic granite.

The K-feldspar megacrystic granodiorite to granite (Unit 4p) underlies a large portion of the Gilbert River belt. The

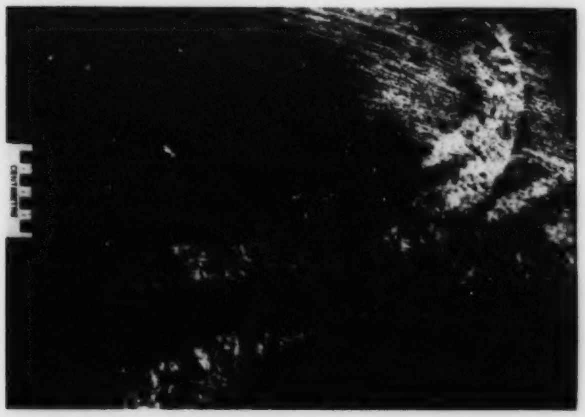

Plate 3. *Well-banded granite gneiss, intruded by discordant pegmatite veins (Gilbert River belt).*

unit occurs as a northwest-trending belt immediately south of the Gilbert River fault, and as an irregularly shaped body south of, and within, the core zone of the Alexis River anorthosite in the northwest. The rock varies from medium to coarse grained, moderately to strongly foliated, pink-white- to red-weathering and having K-feldspar megacrysts comprising 10 to 50 percent of the total rock. Mafic minerals include biotite ± hornblende ± garnet ± muscovite + an opaque phase. The megacrysts are typically recrystallized, although relatively fresh, unrecrystallized grains are present in some areas. Gneissic equivalents were observed as gradations from the foliated megacrystic rock in a few localities. Foliated granodiorite, granite and sillimanite-bearing pelitic gneiss are present as map-scale enclaves.

In contact with the K-feldspar megacrystic unit to the northwest is an extensive body of weak to moderately foliated granite (Unit 4g). The rock is pink- to red-weathering, medium to coarse grained, weak to moderately foliated, with rare diffuse banding. Mafic minerals include biotite ± hornblende ± garnet + an opaque mineral. Although outcrop is poor, the similarity of the texture of this granite, and the matrix texture of K-feldspar megacrystic granite to the southeast, suggest that the two units may be gradations of one another. Small areas of granodiorite and granite gneiss are present in this unit.

A zone of coarse-grained to pegmatitic, pink- to white-weathering, massive to weakly foliated alkali-feldspar granite (Unit 8) occupies the core area of the metasedimentary wedge in the northeast of the belt. The granite is characterized by graphic texture of K-feldspar and quartz, and amphibolite occurs as associated enclaves within. The unit is an extension of a pegmatite field mapped to the north in the Paradise River map region (Gower *et al.*, 1985).

Alexis River Anorthosite (Unit 6)

A distinct unit in the Gilbert River belt is the Alexis River anorthosite (Gower *et al.*, 1987). The unit has a strike length of approximately 150 km. The anorthosite occurs as a 1- to 2-km-wide linear belt extending across the northeast section of the area, and correlates with the unit previously mapped in the Paradise River and Port Hope Simpson map regions (Gower *et al.*, 1985, 1987). Although outcrop is poor, low-magnetic signatures that characterize the unit allow extrapolation across the area.

The dominant rock type of the Alexis River anorthosite in the region is a white-weathering, medium- to very coarse-grained, massive to gneissic, recrystallized anorthosite to leucogabbro–norite. Variable states of recrystallization and strain are evident in some outcrops. The rock ranges from one having a massive, mottled appearance due to patchy distribution of mafic phases to one having a strongly foliated to gneissic, lenticular texture, defined by recrystallized aggregates of mafic phases in a plagioclase-rich matrix. Minerals include plagioclase, amphibole, orthopyroxene, clinopyroxene, biotite and magnetite. Corona reaction textures having orthopyroxene cores rimmed by amphibole and biotite, are locally developed. All phases show variable states of recrystallization, depending on degree of strain. Mafic mineral content varies from less than 1 to 30 percent of the total rock. Pyroxene and amphibole crystals up to 40 cm in length are present in places. Well-developed igneous layering is preserved in some areas, defined by cm-scale alternations of hornblende- and plagioclase-rich layers (Plate 4). At one locality, discordant and concordant, fine- to medium-grained amphibolite dykes were emplaced into the anorthosite.

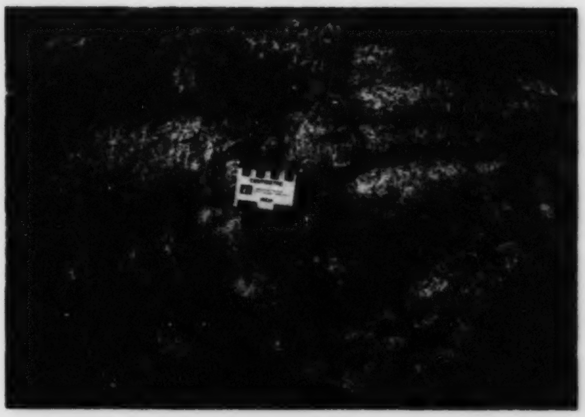

Plate 4. *Igneous layering in Alexis River anorthosite; dark layers consist of hornblende, pyroxene and opaque minerals (Gilbert River belt).*

Associated with the anorthosite is a dioritic to amphibolitic gneiss, which in the eastern portion of the belt flanks both sides of the anorthosite. These rocks are dark-weathering, diffuse to well-banded, medium grained and contain hornblende, biotite and an opaque mineral as mafic phases. Although relationships between the anorthosite and diorite units within the region are not defined, Gower *et al.* (1987) note a similar association in the Port Hope Simpson map region, and suggest that the dioritic gneisses are metamorphic derivatives of anorthosite-related rocks.

MEALY MOUNTAINS TERRANE

The Mealy Mountains terrane is distinguished from the Gilbert River belt by a less intense and more variable regional deformation (except near the northern margin) and more diverse rock types. The terrane is described here in terms of four separate areas, the central portion, the southeast corner, the Upper Paradise River pluton, and the western portion.

Central Portion

The central portion of the Mealy Mountains terrane is underlain by northwest-trending units of variably deformed K-feldspar megacrystic granodiorite to granite, foliated diorite to quartz diorite, granodiorite orthogneiss, psammitic gneisses and foliated granodiorite and granite.

Immediately south of the Gilbert River belt, the boundary with the Mealy Mountains terrane is marked by a northwest-trending fault. The fault is an extension from the Port Hope Simpson region to the west and from the Paradise River region to the north (Gower *et al.*, 1987, 1985), and defines the boundary between the Lake Melville terrane (including the Gilbert River belt) in the northeast and the Mealy Mountains terrane to the southeast. In the southeast of the study area, the fault is marked by breccia, extensive chloritization and hematization and a strong airphoto lineament. In the northwest, a more ductile regime is indicated by the presence of a mylonitic fabric and attenuated and rotated garnets, showing a dextral sense of movement (Plate 5). The presence of brittle and ductile zones suggests that either the northwestern portion represents a deeper section of the fault or that the brittle deformation is a result of a later (Grenvillian ?) reactivation along the fault.

Plate 5. *Rotated garnet porphyroblast in quartz diorite gneiss; implies dextral sense of movement (boundary of Gilbert River belt and Mealy Mountains terrane).*

South of the boundary, a 1- to 5-km-wide zone is underlain by northwest-trending belts of orthogneiss, metasedimentary gneiss and foliated granitoid rocks. In the southeast, a wedge of psammitic gneiss (Unit 2w) is bounded to the north by the fault. The rock is white- to rusty-weathering, medium to coarse grained, layered to diatexitic and locally contains thin (cm to m scale) intercalations of sillimanite-bearing pelitic gneiss. Garnet is common as a restitic phase, along with biotite, and an opaque mineral. Within the psammitic gneiss, km-scale zones of pelitic gneiss and rare calc-silicate rocks are present. The pelitic gneiss is white- to pink-weathering, well banded, and contains sillimanite, biotite, garnet and an opaque phase. A small zone of muscovite-rich pelitic gneiss occurs immediately south of the fault in the east part of the map area. The calc-silicate rock (observed in only two outcrops) is dark-weathering, fine to medium grained, thinly banded, strongly folded and comprises layers containing plagioclase, quartz and minor K-feldspar, alternating with hornblende, and locally clinopyroxene, biotite and opaque mineral-rich layers (Plate 6). Gower *et al.* (1987) noted calc-silicate rocks within pelitic and semipelitic gneiss to the east in the Port Hope Simpson region. A thin zone of foliated granodiorite was also mapped as a continuous belt across the northern portion of the region. The rock is pink- to white-weathering, medium grained, moderate to strongly foliated. The minerals present include biotite, hornblende, minor garnet, and an opaque phase.

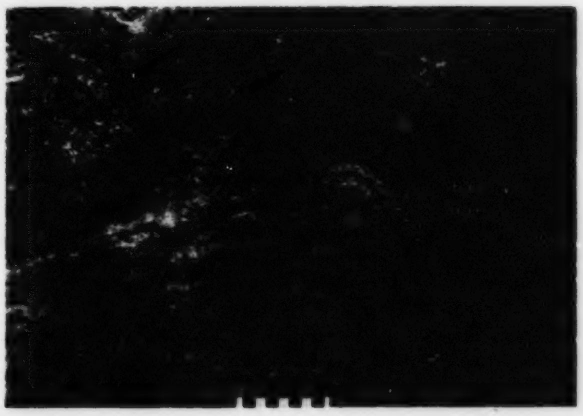

Plate 6. *Thinly banded calc-silicate rock; dark layers consist of hornblende, biotite, pyroxene and opaque minerals (Mealy Mountains terrane).*

South of these rocks are large elongate to oblate-shaped intrusions of K-feldspar megacrystic granodiorite to granite, foliated diorite and foliated granodiorite and granite. These intrusions lack the typical pervasive fabric developed in the rocks to the north, suggesting lower strain southward from the Gilbert River fault.

The K-feldspar megacrystic granite to granodiorite (Unit 4p) is the dominant rock type and occurs as two elongate, northwest-trending intrusions. The map pattern suggests that the intrusions were originally a continuous body and subsequently intruded by diorite, granodiorite and granite, however, no field relations were observed to support this. The rock is red-, pink- to white-weathering, medium to coarse grained, weak to strongly foliated. It has K-feldspar megacrysts ranging from 2 to 10 cm and comprising 10 to

50 percent of the total rock. Mafic minerals include biotite, hornblende, and an opaque mineral, minor garnet and muscovite. The K-feldspar megacrysts vary from being relatively unrecrystallized to showing a 'pseudo-megacrystic' texture. In the latter, megacrysts are strongly recrystallized and grain boundaries are barely recognizable (Plate 7). This texture is characteristic at or near the margins of the intrusions, reflecting more intense deformation (i.e., recrystallization). Attenuated and recrystallized megacrysts exhibit a gneissic appearance in outcrops where the megacrysts reach 10 cm in length. Another characteristic feature of this unit is inhomogeneous strain shown by the rock. Gradations from massive to gneissic texture within single outcrops were observed in several localities (Plate 8). In the northwest part of the larger intrusion, a thin zone of K-feldspar megacrystic granite gneiss was mapped as a separate unit (Unit 1p) The rock is pink- to red-weathering, medium to coarse grained, and is characterized by a strong gneissic fabric, defined by strongly attenuated and recrystallized K-feldspar megacrysts and alternating quartz, plagioclase, feldspar-rich and biotite-rich selvages. These rocks are similar to gneissic gradations observed at several outcrop localities. Map-scale enclaves within this unit include pelitic and psammitic metasedimentary gneiss, granodiorite gneiss and coarse-grained, recrystallized leucogabbro. A thin arcuate zone of dominantly psammitic gneiss occurs within a large body of K-feldspar megacrystic granite. The rock is similar to psammitic gneiss described earlier and contains small enclaves of semipelitic gneiss. One locality of pelitic gneiss contains the assemblage garnet + cordierite + sillimanite + biotite. The presence of cordierite is a diagnostic feature of pelitic gneiss of the Paradise metasedimentary belt in the Hawke River terrane (Gower *et al.*, 1986, 1987), however, it is a rare occurrence in the Mealy Mountains terrane. The cordierite occurs as small (1 to 2 mm) grains, partially replaced by muscovite, biotite, magnetite and quartz. Only a few thin sections of pelitic gneiss have been examined thus far, hence the occurrence of cordierite may be more widespread than that indicated.

Plate 7. *'Pseudo-megacrystic' phase of K-feldspar megacrystic granite (Unit 4p); note diffuse contacts of K-feldspar megacrysts (Mealy Mountains terrane).*

Plate 8. *Inhomogeneous strain in K-feldspar megacrystic granite (Units 4p and 1p, south of the Alexis River, Mealy Mountains terrane).*

A large intrusion of foliated diorite to quartz diorite (Unit 4d) flanking an intrusion of megacrystic granite occurs in the central portion. A homogeneous diorite is the dominant rock type and is grey-, green- to brown-weathering, fine to medium grained, homogeneous to agmatitic, and weak to strongly foliated. The minerals present are plagioclase + minor quartz + hornblende + biotite ± garnet ± pyroxene + an opaque mineral (Plate 9). Hornblende is the dominant mafic phase, constituting up to 40 percent of the total rock. Garnet is common, although not a ubiquitous phase, and crystals up to 5 cm in diameter are locally present near the contact (Plate 10). These grains are typically rimmed by plagioclase + quartz, a texture that suggests a partial replacement of the garnet. Diorite to quartz-diorite gneiss occurs near the margins of the body and probably represents a more deformed equivalent of the foliated diorite. Amphibolite bands are common, in some areas giving the rock a gneissic appearance, probably representing deformed mafic dykes. At a few localities, orthopyroxene and clinopyroxene are stable in a strongly recrystallized phase of the diorite, an association that is suggestive of granulite-facies conditions. Abundant enclaves and/or minor intrusions of granodiorite, K-feldspar megacrystic granite, metasedimentary gneiss, granodiorite gneiss, and minor leucogabbro are present.

The intrusion is a lithological anomaly within the region, being of more calcic-alkaline composition compared with the dominant potassic-rich granitoids.

Southeast Corner

The southeast corner of the region consists dominantly of interfingering K-feldspar megacrystic granodiorite to granite and pelitic to psammitic metasedimentary gneisses.

The pelitic gneiss (Unit 2s) is cream, buff- to rusty-weathering, medium grained and characterized by schlieric to layered migmatitic texture (Plate 11). The rock comprises a K-feldspar, plagioclase, quartz leucosome, and sillimanite,

Plate 9. *Strongly foliated, homogeneous diorite (Unit 4d; Mealy Mountains terrane).*

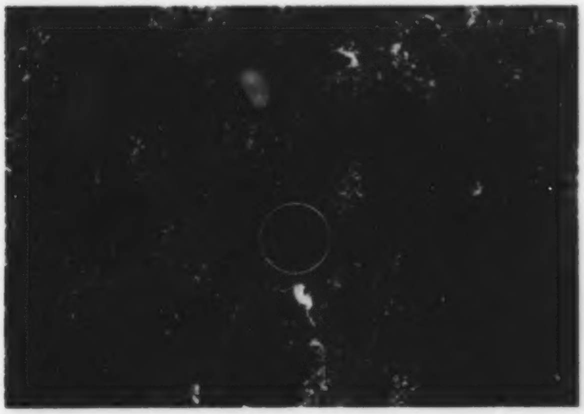

Plate 10. *Large garnet porphyroblasts in foliated diorite (Unit 4d). Note plagioclase + quartz reaction rims around garnets. (Near contact of diorite intrusion, Mealy Mountains terrane).*

Plate 11. *Well-banded, strongly folded sillimanite-bearing metasedimentary gneiss (Unit 2s; Mealy Mountains terrane).*

garnet, biotite, muscovite and an opaque mineral in the restite. Sillimanite is widespread, and locally occurs as aggregates in layers up to 2 cm wide. Garnet is a near ubiquitous phase and occurs as euhedral grains up to 3 cm in diameter. In some areas, a white-weathering diatexitic component is associated with the pelitic gneiss. The rock is typically homogeneous, although irregular selvages of biotite + garnet + an opaque mineral are common. Garnet is a common mineral, occurring both in the quartz + feldspar-rich zones and in the restite component. Granite and pegmatite veins are normally associated with this rock type. Mappable units of psammitic gneiss as gradations from the pelitic gneiss were distinguished. The psammite (Unit 2w) is rusty-, white- to buff-weathering, medium to coarse grained, and ranges from well banded to diatexitic. Minerals present include K-feldspar, plagioclase, quartz, biotite, garnet, and an opaque mineral and locally sillimanite and muscovite. Thin (cm scale), intercalated layers of sillimanite-bearing pelite occur in some areas. Concordant granite and alkali feldspar granite veins are common and in many areas, enhance the overall gneissic appearance (Plate 12).

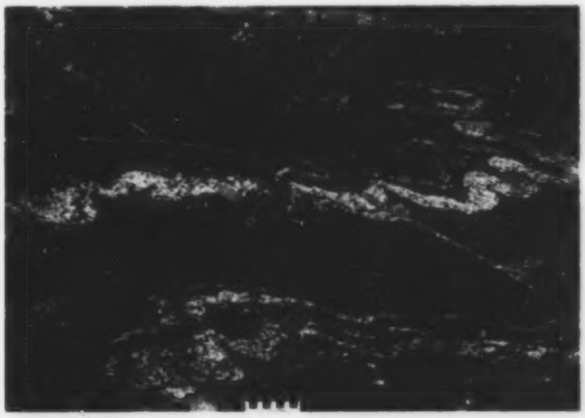

Plate 12. *Well-banded psammitic metasedimentary gneiss (Unit 2w). Intruded by concordant and discordant granite veins; south of the Alexis River, Mealy Mountains terrane.*

Associated with both the pelitic and psammitic gneiss are small (m scale) intrusions of a white-weathering, garnet-bearing alkali feldspar granite. The consistent association with metasediments and the presence of garnet implies a derivation from the sediments.

The K-feldspar megacrystic granitoid rocks (Unit 4p) are pink- to white-weathering, medium to coarse grained, and weak to strongly foliated (rarely gneissic). These occur as irregular-shaped bodies intruding the host metasediments. Biotite, garnet and hornblende are the dominant mafic phases. Garnet is near ubiquitous in these rocks in contrast to megacrystic rocks in other parts of the region. This is interpreted to be a result of assimilation of the granite and the host, garnet-bearing metasediments. In some areas (notably near the contact), a 'pseudo-megacrystic' texture is developed where grain boundaries of recrystallized K-feldspar megacrysts are diffuse and partially incorporated into the matrix component of the rock. Contacts with the sediments are typically diffuse and in some areas being gradational over hundreds of metres. Such a relationship implies extensive mixing and assimilation of the units, probably at considerable depth. Strain gradients within the megacrystic unit are highly variable and gradations from massive to gneissic are locally developed at outcrop scale.

Both the metasediments and the K-feldspar megacrystic granitoid rocks contain abundant enclaves and/or intrusions (m to km scale) of the following rock types: (i) foliated hornblende + biotite diorite, which is related to the large intrusion in the central portion, (ii) foliated, sugary-textured biotite granite to granodiorite, a common intrusive phase into the K-feldspar megacrystic granite to granodiorite and, (iii) a small body of leucogabbro. Noteworthy, is a small intrusion of muscovite-rich granite (Unit 4u) along the eastern edge of the area. The rock is distinct from other foliated granites in that it contains up to 20 percent muscovite. It is pink-weathering, medium grained, moderate to strongly foliated and contains abundant partially assimilated sillimanite-bearing pelitic gneiss enclaves. Gower *et al.* (1988a) used the term the Southwest Pond domain for a zone of K-feldspar megacrystic granitoid rocks, and metasedimentary gneiss, in the northwest corner of the St. Lewis map area. At the time, inclusion of this area in the Mealy Mountains terrane was withheld pending further mapping. Subsequent mapping in the Alexis River region now allows the domain to be included as part of the Mealy Mountains terrane, and therefore, the name Southwest Pond domain should be abandoned.

Upper Paradise River Pluton

The western area is underlain largely by a lobate-shaped intrusive body, informally named the Upper Paradise River pluton. The body consists of syenitic to quartz monzonitic rocks, having associated gabbro norite and minor granite enclaves. Outcrop distribution in this area is relatively sparse and relationships between units are poorly exposed. The rocks exhibit variable degrees of deformation and recrystallization and are characterized by a high aeromagnetic anomaly. Eade (1962) reported outcrops of monzonitic to granodioritic rocks within what have now been mapped as the margins of the body.

The intrusion has been subdivided into three major rock types, namely monzonite (Unit 4m); syenite to quartz syenite (Unit 4y); and K-feldspar megacrystic monzonite to quartz monzonite (Unit 4mp).

The dominant rock type underlying the central portion of the pluton is a brown- to grey-weathering, moderately to strongly foliated, medium- to coarse-grained, variably recrystallized monzonite. Minerals include biotite + hornblende ± clinopyroxene + zircon + titanite. Gradations to quartz monzonite on an outcrop scale are common.

The western margin of the pluton is underlain by a brown, pink- to white-weathering, medium- to coarse-grained, moderate to strongly foliated, recrystallized alkali-feldspar syenite to quartz syenite, containing hornblende + biotite + clinopyroxene + an opaque mineral. In some areas, a sub-gneissic fabric is developed, defined by thin (0.25 cm) quartz-rich lenses parallel to main fabric. Rocks within this unit, although showing variable fabric intensity, are consistently strongly recrystallized. Mafic mineral content is also variable, but the similar bulk-rock composition of the rocks suggests they are part of the same intrusion.

A distinctive and homogeneous subunit within the Upper Paradise River pluton, is a K-feldspar megacrystic monzonite to quartz monzonite. The rock is honey-brown weathering, medium to coarse grained, massive to weakly foliated and characterized by the presence of grey-weathering K-feldspar megacrysts, constituting 10 to 30 percent of the total rock (Plate 13). The K-feldspar megacrysts are generally unrecrystallized, although in some areas, a thin rim of polygonal recrystallized grains is developed. Minerals include plagioclase, quartz, K-feldspar, hornblende, biotite clinopyroxene, and an opaque phase. Enclaves and/or intrusions of foliated monzonite, diorite and monzonite gneiss are present.

Plate 13. *Weakly foliated to massive K-feldspar megacrystic monzonite (Unit 4mp). K-feldspars locally have recrystallized rims; Mealy Mountains terrane.*

In the southeast corner, two apparently isolated intrusions of monzonite were mapped. Although these intrusions occur exterior to the main monzonite body, they are identical, both texturally and mineralogically, to rocks within the main body, and thus, are considered a part of the Upper Paradise River pluton. One of the intrusions consists of foliated monzonite having core zones of a K-feldspar megacrystic phase. This may represent a more slowly cooled portion of the exterior recrystallized monzonite.

Other units, mapped as variants of the three major units include a K-feldspar megacrystic alkali feldspar syenite (Unit 4yp) and a monzonitic gneiss (1m). The syenitic to quartz-syenitic rock occurs as a small body within the foliated monzonite unit and as a separate intrusion flanking the southeast edge of the Upper Paradise River pluton. Similarities between these rocks and the three main rock types suggest that they are part of the Upper Paradise River pluton. The rock is brown- to grey-weathering, medium to coarse grained and characterized by a strong foliation and recrystallized K-feldspar augen. Hornblende, clinopyroxene, minor biotite and an opaque mineral are the mafic phases. An elongate body of monzonite gneiss (Unit 1m) occurs within the K-feldspar megacrystic monzonite unit of the Upper Paradise River pluton. A single outcrop of this rock also occurs along the Paradise River, although its limited extent prevents it from being depicted on the map. The rock is brown- to grey-weathering, medium grained, strongly recrystallized and well banded. The gneissosity is defined by alternations of cm-scale mafic-rich and plagioclase + quartz-rich bands. The isolated occurrence and the limited extent of these rocks suggest that they are strongly deformed enclaves within a less-deformed host rock. Associated with the three main units are map-scale bodies (intrusions ?) of gabbro and gabbronorite, which typically underlie small topographic highs in the area. The rocks are mineralogically and texturally similar to gabbroic units in the northeast corner of the region, within the White Bear Arm complex. Gower *et al.* (1987) noted a similar association of monzonitic, granitic and gabbronoritic rocks in the White Bear Arm complex. They observed layered units showing transitions from gabbronorite to monzogabbro to monzonite, as well as enclaves of gabbronorite in monzonite and suggested that several periods of gabbronorite and monzonite emplacement have occurred. Although relationships between the gabbro and monzonite units are not exposed in the Alexis River region, a similar interpretation may be applicable here.

Speculation can be made regarding age relations between these units, based on their degree of deformation and km-scale enclaves or intrusions of the K-feldspar megacrystic monzonite and foliated monzonite units in the Upper Paradise River pluton. The pervasively strong foliation and extensive degree of recrystallization present in the syenite to quartz-syenite unit suggest that it is older than the foliated monzonite and the weakly deformed, K-feldspar megacrystic units. Similarly, the monzonite unit would appear to be older than the megacrystic phase. Both the foliated monzonite and the megacrystic phase of the pluton occur within the strongly foliated syenite as km-sized bodies. However, contacts between these rocks were not observed and to choose between enclaves or small intrusions would only be speculative. In one locality, in the northeast portion of Upper Paradise River pluton, a small enclave of strongly foliated quartz syenite was found hosted by the K-feldspar megacrystic unit. Diffuse contacts of the enclave suggest that the two rocks types may be of similar age. Intended U–Pb radiometric dating of samples collected from the Upper Paradise River pluton will presumably establish age relations and time of emplacement of the various intrusive units.

In the St. Lewis River Map region (southeast of the Alexis River region), the Pinware terrane was defined by Gower *et al.* (1988a). The Pinware terrane is a distinct zonation in the Grenville Province in southeastern Labrador composed principally of granitoid plutons of pre-Grenvillian and Grenvillian ages (Gower and Loveridge, 1987). Grenvillian and post-Grenvillian plutons have syenite to monzonite to alkali-feldspar-granite compositions and are characterized by a donut shaped aeromagnetic plateau having high rims and low interiors. Although the Pinware terrane has not been mapped west of the St. Lewis River map region, Gower *et al.* (1988a) have extrapolated the northern limit of the terrane to just south of the Alexis River region (on the basis of the characteristic circular aeromagnetic anomalies).

A similar, although less well-pronounced anomaly, is characteristic of one of the satellite intrusions of the K-feldspar megacrystic monzonite. High aeromagnetic anomalies also characterize the Upper Paradise River pluton as a whole. Signatures, however, are variable, and not well defined, presumably due to the composite nature of the intrusion. The characteristics of the K-feldspar megacrystic monzonite unit (low degree of strain and recrystallization, and aeromagnetic signature) invites the possibility of the rocks having a Grenvillian or late syn-Grenvillian age.

Western Portion

In the northwest, flanking the northern arm of the Upper Paradise River pluton, are bodies of amphibolitic (Unit 1a) to dioritic orthogneiss (Unit 1d), granodiorite gneiss (Unit 1b), hornblende-rich, strongly recrystallized syenite to quartz syenite (Unit 4h), and foliated granite (Unit 4g). Characteristic of the units in this area is a pervasive, north-northwest to north-trending fabric, which is distinct from the typical northwest regional trend. Amphibolitic to dioritic gneisses underlie an open arcuate belt. These rocks are black- to grey-weathering, fine to medium grained, and typically well banded. Hornblende is the dominant mafic phase comprising up to 60 percent of the total rock.

Enclaves within the gneisses include leucogabbro, granite and a west-trending enclave of sillimanite-bearing metasediment. The locality of this outcrop of metasediments is anomalous in the region in that it is isolated from similar rocks that are associated with K-feldspar megacrystic granitoid rocks farther to the northeast.

Flanking the western edge of the Upper Paradise River pluton is a north-trending belt of granodiorite to granite gneiss. These rocks are medium grained, diffusely to well banded, and recrystallized. The gneissosity is concordant with the western margin of the intrusion, suggesting that the trend of the gneisses is at least partly a result of emplacement of the later body. Enclaves hosted in the gneisses include foliated diorite to diorite gneiss.

The extreme northwest corner of the region is underlain by the extension of a hornblende-bearing intrusion in the Paradise River map region (Gower *et al.*, 1985), where it was mapped as a granodiorite to monzonite. In the Alexis River region, syenitic to quartz syenitic rocks dominate. The rock is cream- to rusty-brown-weathering, medium grained, and characterized by a strong, recrystallized fabric. Ubiquitous hornblende is the dominant mafic phase, biotite, clinopyroxene, orthopyroxene and an opaque phase are also present.

Enclaves and/or intrusions in this unit include; metagabbro, foliated granite and minor granite gneiss. The metagabbro is a distinct rock that is black-weathering, coarse grained, massive, and intruded by abundant pegmatitic and granite veins. Minerals include plagioclase, opaque, and hornblende pseudomorphs of clinopyroxene. Coarse pegmatite veins intruding the gabbro are strongly deformed and sheared, locally showing gradation from undeformed pegmatite to a mylonite at the immediate contact with the gabbro, although the gabbro retains an apparently massive texture. Small veins of recrystallized granite to quartz syenite intruding the gabbro may be related to the host syenite, suggesting remelting during gabbro emplacement.

In the extreme southwestern part of the Paradise River map region, this unit has been dated at 1735 and 1718 Ma from U–Pb determinations on two zircon fractions, and 1631 ± 1 Ma from two monazite fractions (Scharer and Gower, 1988). As the rocks show no indication of a high-grade overprint that would explain the new growth of monazite, they concluded that the monazite dates the time of rock crystallization and the zircons, grown coevally with monazite, include inherited older components. A $^{40}Ar/^{39}Ar$ plateau age of 980 ± 3 Ma was obtained from hornblende from the same rock (van Nostrand, 1988). The age is interpreted as a Grenvillian cooling, below approximately 550°C.

POST-GRENVILLIAN MAFIC DYKES

Two suites of post-Grenvillian mafic dykes were mapped within the Alexis River map region. The first suite includes 2 separate northeast-trending gabbro dykes, which are extensions of dykes mapped farther to the north in the Paradise River and Sand Hill River map regions (Gower *et al.*, 1985, 1986). Exposure is limited to a single outcrop for one of the dykes in the west and 3 outcrops for the other in the eastern portion. The rock is typically a brown-weathering, massive, coarse grained, ophitic to sub-ophitic textured gabbro. Plagioclase, clinopyroxene, opaque and olivine are the mineral phases present. Characteristic interstitial K-feldspar, present in all localities, varies from less than 5 percent in the eastern dyke to 15 percent in the western dyke. The high proportion of K-feldspar in the western dyke is matched only by that present in the same dyke in the Paradise River map area. The variation in K-feldspar content may reflect the relative depths of intrusion, being concentrated in the fractionated top part of the dykes.

The western most dyke is on strike with a 100-km-long dyke, extending from north of Cape Porcupine to the southwest corner of the Paradise River region (Gower *et al.*, 1985). Extrapolation into the Alexis River region continues the strike length of the dyke to approximately 150 km. A strong airphoto lineament in the southwest of the area is on strike with the exposed dyke on the Paradise River, and probably extends the dyke to the southwest corner of the map area. The dyke in the eastern part of the area, extrapolated to the northeast, links up with a gabbro dyke in the Sand Hill River region to have a strike length of at least 100 km.

The second suite of dykes mapped are brown- to black-weathering, fine grained, undeformed, massive and show sharp chilled margins with the country rock (Plate 14). The dykes were observed only in two localities. A 20- to 60-cm-wide, 060°-trending dyke intrudes a foliated diorite along the Paradise River in the west and a single similar trending dyke postdates a granite gneiss in the east of the region. In thin section, the rock comprises a diabasic texture defined by fresh, euhedral plagioclase laths with intergranular clinopyroxene, olivine, opaque and biotite. Very local alteration is evidenced by replacement of mafic phases by epidote. Similar dykes have been mapped in the Port Hope Simpson area, in the southern parts of Sandwich Bay. A K–Ar whole-rock date of 327 ± 13 Ma for one dyke in Sandwich Bay was reported by Murthy *et al.* (1989).

Plate 14. *Fine-grained, post-Grenvillian diabase dyke (Unit 12), intruding foliated quartz diorite (Unit 4d); Upper Paradise River, Mealy Mountains terrane.*

STRUCTURE

The regional structure of the Alexis River map region is dominated by a pervasive northwest trend in the northeastern area and a variable west to northwest trend in the southern and northwestern areas. The trends are developed in all units with the exception of post-Grenvillian dykes. Regional fabric trends within the region are indicated on Figure 3.

In the White Bear Arm complex, fabrics are limited to a weak foliation in metagabbro and gneissosity in paragneiss and granite enclaves. Granite enclaves within the gabbro are characterized by a strong, submylonitic fabric. In one outcrop, a well-developed C–S fabric suggests a east over west (dextral) sense of movement (Plate 1). This is consistent with the regional dextral movement proposed by Gower *et al.* (1987) and Hanmer and Scott (1990). Although no outcrop was mapped along the Gilbert River Fault (Gower *et al.*,

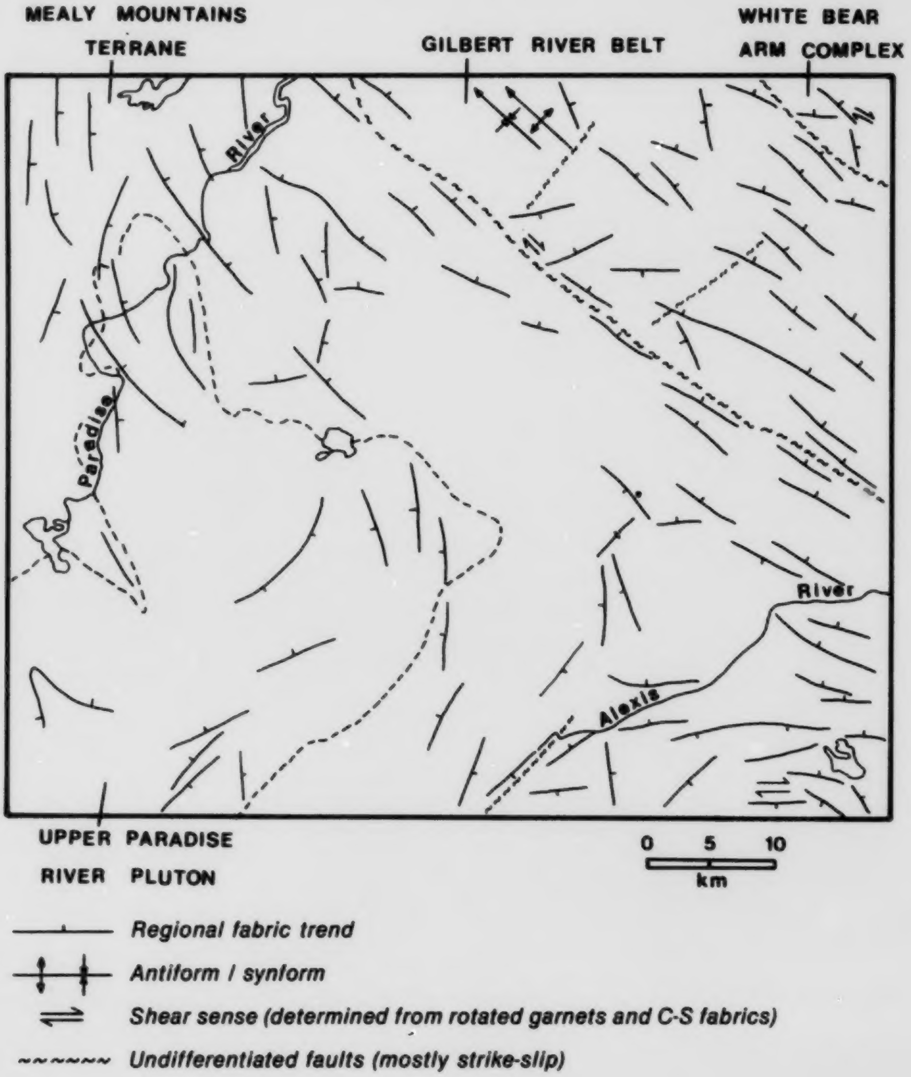
MEALY MOUNTAINS TERRANE
GILBERT RIVER BELT
WHITE BEAR ARM COMPLEX
River
Paradise
River
Alexis
UPPER PARADISE RIVER PLUTON
0 5 10
km
Regional fabric trend
Antiform / synform
Shear sense (determined from rotated garnets and C-S fabrics)
Undifferentiated faults (mostly strike-slip)

Figure 3. *Major structural features and lithostructural subdivisions in the Alexis River map region.*

1987), a strong airphoto and aeromagnetic signature and extrapolation from adjacent map regions, defines the fault.

The Gilbert River belt is characterized by a pervasive, northwest-trending foliation and gneissosity. The northwest portion of the Alexis River anorthosite, is interpreted on the map as a northwest-plunging antiformal/synformal structure. Mapping by Gower *et al.* (1985) to the north and aeromagnetic signatures suggest that the southern limb of the anorthosite has thickened as a result of folding about northwest-plunging axes. Fabrics in the Mealy Mountains terrane show a regional trend varying from northwesterly, near the boundary with the Gilbert River shear belt, west to northwest in the southeast corner of the region to a northward curvature of the fabric in the west and northwest areas.

Fold styles are typically complex, ranging from open to tight Z and M folds in orthogneiss and granitoid rocks to isoclinal and typically transposed in metasedimentary gneiss (Plates 15 and 16). Fold axes generally show a shallow to moderate plunge toward the northwest, although locally trend steeply toward the northeast. Mineral lineation and rodding, measured on foliation planes typically plunge moderately toward the northwest in the Gilbert River shear belt and vary from a shallow to moderate plunge toward the northwest to north in the Mealy Mountains terrane.

Kinematic indicators, in particular rotated porphyroblasts, strong lineations, and rare C–S fabrics found along or near contacts, suggest a dextral shear sense (Plate 5), and are indicated on Figure 3. Attenuated garnets in orthogneisses along the southern boundary of the Gilbert River shear belt show aspect ratios up to 15:1. Gower *et al.*

Plate 15. *Z-folds in well-banded metasedimentary gneiss; Mealy Mountains terrane.*

Plate 16. *Open M-folds in well-banded granite gneiss; Banding enhanced by concordant granite veins; Gilbert River belt.*

(1987) proposed a regional kinematic framework in which the Lake Melville terrane (including the Gilbert River belt) has moved dextrally with respect to the Mealy Mountains terrane. Hanmer and Scott (1990) suggested a similar sense of movement within the Gilbert River belt, on the coast of Labrador. Late (post-Grenvillian) northeast-trending, brittle faults, are marked by extensive hematization, chloritization, quartz veining and fracturing. Amount of displacement (if any) along these faults is undetermined. In the northwest corner of the map region, a 5- to 10-m-wide quartz vein, trending 030 to 050° occurs along the Paradise River. The quartz vein is on strike with the Paradise River fault to the north (Gower *et al.*, 1985). The Paradise River fault is interpreted to define (in part) the Sandwich Bay graben, and is probably of similar age to the Double Mer and Lake Melville grabens farther northwest (Gower *et al.*, 1983).

METAMORPHISM

Metamorphic grade in the region ranges from middle to upper amphibolite facies with evidence that granulite facies was locally attained. Garnet is a common metamorphic mineral in metasedimentary gneiss, orthogneiss, granitoid rocks and associated mafic rocks. Variable degrees of partial melting are present in most units, ranging from small (cm scale) segregations in granitoids, orthogneiss and mafic rocks to extensive, and locally complete anatexis in metasediments. The presence of metamorphic orthopyroxene in a few localities suggests that granulite-facies conditions were reached.

Metasedimentary gneiss is characterized by the assemblage biotite + sillimanite + garnet + K-feldspar + plagioclase + quartz, with magnetite as a common accessory phase. Garnet and biotite are commonly found in the leucosome component. One locality of pelitic gneiss in the Mealy Mountains terrane contains cordierite + garnet + sillimanite + biotite. Some of the cordierite porphyroblasts are partially retrogressed according to the reaction:

cordierite + K-feldspar = sillimanite + quartz ± biotite

Other grains are partially to completely pinitized, having no evidence of replacement by the above phases. Gower *et al.* (1987) note that cordierite is a widespread metamorphic mineral in the Paradise metasedimentary gneiss belt but was not seen in the Gilbert River belt or the metasedimentary gneiss in the southwest area of the Port Hope Simpson region (Mealy Mountains terrane).

Granitoid rocks and orthogneiss contain biotite ± hornblende ± garnet ± clinopyroxene + an opaque mineral as constituent mafic minerals. Recrystallized clinopyroxene is present in syenitic and monzonitic rocks of the Upper Paradise River monzonite. Retrogression to lower or middle amphibolite facies is indicated by the presence of post- to syn-fabric muscovite in metasedimentary gneiss and orthogneiss, and replacement of hornblende and/or clinopyroxene by chlorite, biotite and epidote in granitoid and mafic rocks.

Mafic rocks of the Alexis River anorthosite and the White Bear Arm complex partially preserve primary (igneous) minerals, but commonly show hydration and replacement of pyroxene by amphibole and recrystallization of plagioclase. Coronitic textures are developed in some areas in gabbroic rocks as olivine cores, mantled by orthopyroxene and amphibole rims. Small bodies of granulite facies, two pyroxene-bearing gabbronorite and diorite occur in the White Bear Arm complex and the Mealy Mountains terrane, respectively.

ECONOMIC POTENTIAL

The economic mineral potential of the Alexis River Region is limited to, (i) local rusty-weathering zones associated with metasedimentary gneiss, (ii) minor pyrite found within late (post-Grenvillian) brittle fault zones in the eastern portion of the region, (iii) trace amounts of pyrite occurring in a Paleozoic dyke in the western portion of the area, and (iv) local anomalous scintillometer readings (200 to 400 cps) determined on few of the pegmatites. Noteworthy are large magnetite crystals (up to 6 cm in diameter) associated with a late alkali-feldspar granite pegmatite intruding a metagabbro enclave just north of the Alexis River.

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GEOLOGICAL MAPPING IN THE TORNGAT OROGEN, NORTHERNMOST LABRADOR: PRELIMINARY RESULTS

R.J. Wardle, M.J. Van Kranendonk[1], F. Mengel[2] and D. Scott[3]
Labrador Mapping Section

ABSTRACT

Northernmost Labrador contains the boundary between the Archean Nain Province and the enigmatic Burwell Terrane of the Early Proterozoic Torngat Orogen. The Nain Province crust consists predominantly of granulite-facies tonalitic to granodioritic gneisses that were extensively migmatized prior to intrusion of an Early Proterozoic mafic dyke swarm and subsequent reworking within the Torngat Orogen. A major feature of the western part of the Nain crust is an approximately 70-km-long Archean anorthosite that strikes north-south, parallel to the main Proterozoic trends in the area. In contrast, the Burwell Terrane consists largely of Early Proterozoic metaplutonic rocks, ranging in composition from diorite to hypersthene-granodiorite and granite, interspersed with belts of metasedimentary gneiss. Metaplutonic rocks of Burwell Terrane are largely in sheared contact with the western limit of Nain crust, but some are seen to have also intruded this crust. The Burwell Terrane is therefore interpreted to represent the deep levels of a magmatic arc that was, in part, constructed upon the western edge of Nain Province, and/or accreted late in the arc's magmatic history.

Proterozoic ductile deformation was associated with amphibolite-facies reworking of the Archean crust and amphibolite-to granulite-facies metamorphism of the Burwell Terrane. Ductile shearing was concentrated in the Nain crust along the margins of the anorthosite body and along the Nain–Burwell contact. Early shearing and folding occurred at granulite to amphibolite facies in an oblique collisional environment, through sinistral shear associated with east-side-up motion. This was followed by a separate phase of ductile shearing and ultramylonite formation in response to west-side-up thrusting. The shear zones located along the boundary of the Nain crust and Burwell Terrane do not represent the type of fundamental plate boundary predicted by previous interpretations.

Concordant zones of sulphide (predominantly pyrite) mineralization have been discovered within Archean anorthosite and at the contact of metasedimentary gneiss with hypersthene-granodiorite in the interior of the Burwell Terrane. These may provide new exploration targets in northern Labrador and adjacent Northwest Territories.

INTRODUCTION

The Early Proterozoic Torngat Orogen (1.86–1.79 Ga) forms the collision zone between the Archean blocks of the Rae Province, in the west, and the Nain Province to the east (Hoffman, 1988, 1990a, b; Wardle *et al.*, 1990a, b) (Figure 1). The Rae Province was extensively metamorphosed and deformed during this collision whereas only the western edge of the Nain Province was similarly affected. Along much of its length in northern Labrador, the core of the collision zone is represented by a belt of intensely deformed and migmatized metasedimentary gneiss known as the Tasiuyak gneiss, which is approximately coincident with the Abloviak shear zone (Van Kranendonk, 1990; Ermanovics and Van Kranendonk, 1990; Van Kranendonk and Ermanovics, 1990; Mengel *et al.*, 1991) (Figure 1). Evolution of the suture was dominated by early oblique collision ca. 1.86 Ga followed by transcurrent, sinistral shearing and late cross-orogen shortening in the interval 1.84–1.79 Ga (Bertrand *et al.*, 1990). The Abloviak shear zone and associated Tasiuyak gneiss can be traced aeromagnetically across Hudson Strait where they establish continuity of the Torngat Orogen into southern Baffin Island (Hoffman, 1990a). In northernmost Labrador, the picture is

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[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, K1A 0E8

[2] Geologisk Museum, Købnhavn Universitet, Oster Voldgade 5-7, 1350 Købnhavn K, Denmark

[3] GEOTOP, Université du Québec, C.P. 8888, Succ. A., Montréal, Québec, H3C 3P8

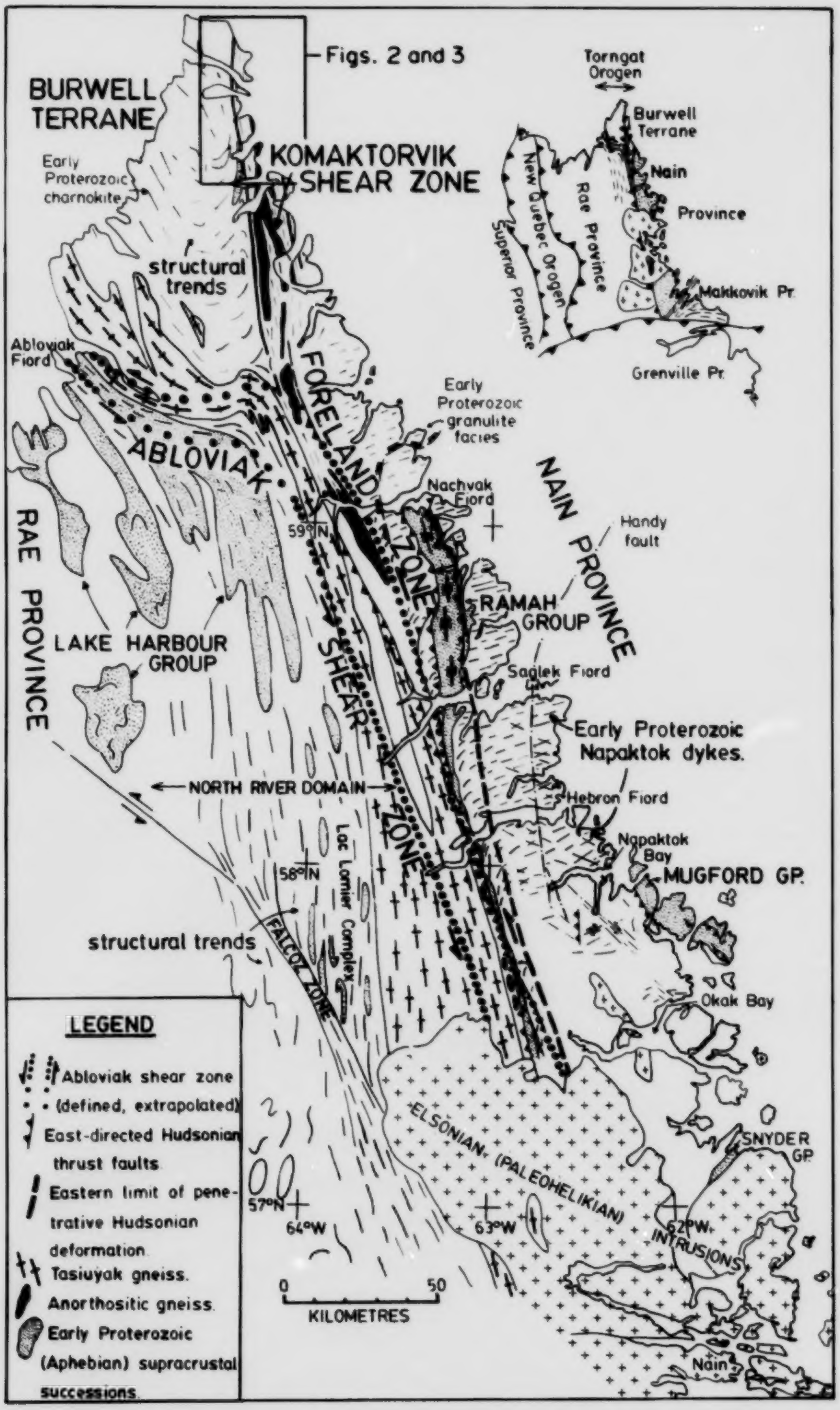
Figs. 2 and 3
BURWELL TERRANE
Early Proterozoic charnokite
KOMAKTORVIK SHEAR ZONE
structural trends
Abloviak Fiord
FORELAND ZONE
ABLOVIAK SHEAR ZONE
Early Proterozoic granulite facies
Nachvak Fiord
NAIN PROVINCE
Handy fault
RAMAH GROUP
RAE PROVINCE
LAKE HARBOUR GROUP
Saglek Fiord
Early Proterozoic Napaktok dykes.
NORTH RIVER DOMAIN
Hebron Fiord
Napaktok Bay
MUGFORD GP.
59°N
58°N
57°N
64°W
63°W
62°W
Lac Lomier Complex
structural trends
FALCOZ ZONE
Okak Bay
ELSONIAN (PALEOHELIKIAN) INTRUSIONS
SNYDER GP.
Nain
0
50
KILOMETRES
Torngat Orogen
Burwell Terrane
Nain Province
Makkovik Pr.
Rae Province
New Quebec Orogen
Superior Province
Grenville Pr.
LEGEND
Abloviak shear zone (defined, extrapolated)
East-directed Hudsonian thrust faults
Eastern limit of penetrative Hudsonian deformation
Tasiuyak gneiss.
Anorthositic gneiss.
Early Proterozoic (Aphebian) supracrustal successions.

Figure 1. *Generalized geological map of Torngat Orogen in northern Labrador, northeastern Québec and Northwest Territories (Van Kranendonk, 1990). Map area discussed in this report is outlined by box labelled Figures 2 and 3.*

complicated by a second aeromagnetic lineament, termed the Komaktorvik zone by Korstgaard *et al.* (1987), or Komaktorvik shear zone (Wardle *et al.*, 1990b), which diverges from the Abloviak zone and isolates an intervening crustal wedge referred to as the Burwell Terrane (Figure 1). Korstgaard *et al.* (1987) noted the lack of information concerning this terrane, but suggested that it was a detached segment of Nain crust on the basis of its aeromagnetic pattern. Hoffman (1990a, b) has more recently proposed the Burwell Terrane as an independent plate with extensions into south Baffin Island and West Greenland. Further interpretation has been hindered by the lack of reliable dates and paucity of geological information for Burwell Terrane; the terrane having been last described during reconnaissance mapping in the early nineteen seventies (Taylor, 1977).

In addition to its scientific importance the region contains Archean anorthosite intrusions that may have potential for mineralization (Wilton and Wardle, 1990; Swinden *et al.*, 1991) and has recently become accessible through a program of northern radar site construction. Studies of the Burwell Terrane, and its boundaries with the Nain and Rae provinces, are therefore a priority target for renewed mapping in northernmost Labrador.

A joint project between the Newfoundland Geological Survey Branch and the Geological Survey of Canada will eventually map all of Labrador north of 59°15'. This report covers mapping in parts of NTS 25 A/1, 2, 7 and 8 (Figure 2); associated mapping to the south in NTS 24/P16 has been reported separately by Van Kranendonk and Scott (1992). Field work for 1991 was entirely boat-supported and restricted to areas accessible from the sea. Inland areas shown as unmapped on the accompanying map (Figure 2) will be completed in 1992 when extensive helicopter support becomes available.

Northernmost Labrador has received cursory visits by geologists engaged in coastal reconnaissance since the last century, e.g., Low (1896), Coleman (1917) and Forbes and Odell (1938). Early mineral-exploration programs by BRINEX (Piloski, 1959) led to a broader understanding, but complete geological coverage (at 1:250 000 scale) was not achieved until the early nineteen seventies (Taylor, 1977, 1979). The current study was preceded by a brief reconnaissance visit in 1990 (Wilton and Wardle, 1990).

The map area, which lies at the northern end of the Torngat Mountain chain, is rugged, mountainous and indented by spectacular fiords. Relief becomes progressively subdued to the west where the land slopes down to the tundra plain surrounding Ungava Bay.

GENERAL GEOLOGY

The area comprises two principal groups of rocks (Figures 2 and 3). In the east, is an Archean gneiss complex, intruded by Early Proterozoic mafic dykes, and variably affected by Early Proterozoic deformation. These rocks are clearly correlative with those of the Nain Province and establish the continuity of Archean crust to the northern tip of Labrador, contrary to previous mapping (Taylor, 1977) that assigned an Early Proterozoic age to all of the area dealt with in this report. The dykes serve as an important chronological marker for distinguishing the effects of Archean versus Proterozoic tectonism. In the western part of the area, metaplutonic rocks and associated metasedimentary gneisses of inferred Early Proterozoic age form the Burwell Terrane.

The map area has been further subdivided into five lithotectonic zones (Figure 3). Extrapolation of these zones through unmapped or water-covered areas has been greatly facilitated by the use of unpublished, colour-shaded-relief aeromagnetic maps produced by G. Kilfoil of the Geological Survey Branch.

Zone 1 comprises Archean, granulite-facies granitoid gneisses interspersed with narrow belts of supracrustal gneiss and mafic granulite, and intruded by a northeast-trending swarm of Early Proterozoic metadiabase dykes. This zone has substantially escaped significant penetrative Proterozoic deformation.

Zone 2 is underlain by Archean gneisses and Proterozoic dykes of similar origin to those in Zone 1, but transected by belts of Early Proterozoic, amphibolite-facies ductile shearing.

Zone 3 consists entirely of Archean anorthosite and related rocks, which are overprinted by Early Proterozoic amphibolite-facies fabrics.

Zone 4 consists of Archean gneisses that preserve widespread Archean granulite-facies assemblages but have been partially retrogressed to Proterozoic amphibolite facies. In contrast to Zone 2, these have not been significantly affected by shearing, except along the boundaries of the zone.

Zone 5 represents the eastern part of the Burwell Terrane and is composed of a western hypersthene-granodiorite batholith and an eastern metaplutonic suite of diorite–quartz diorite–tonalite and granodiorite composition, interspersed with belts of supracrustal gneiss.

Late intrusions of metaplutonic rocks have been found in Zones 2 to 5 with the greatest concentration being in Zone 5 (Burwell Terrane). Van Kranendonk and Scott (1992) have also identified similar rocks in the equivalents of Zone 1 to the south. The metaplutonic rocks are probably largely or entirely of Early Proterozoic age; however, in the absence of geochronological data a Late Archean age cannot be ruled out for some of them.

As defined by Stockwell (1982) the term 'Nain Province' refers only to Archean crust unaffected by Proterozoic tectonism, and would exclude most of the Archean rocks of the map area. In this report, the term 'Nain crust' is used to refer to such rocks of Nain Province affinity, regardless of the degree to which they have been affected by Proterozoic deformation.

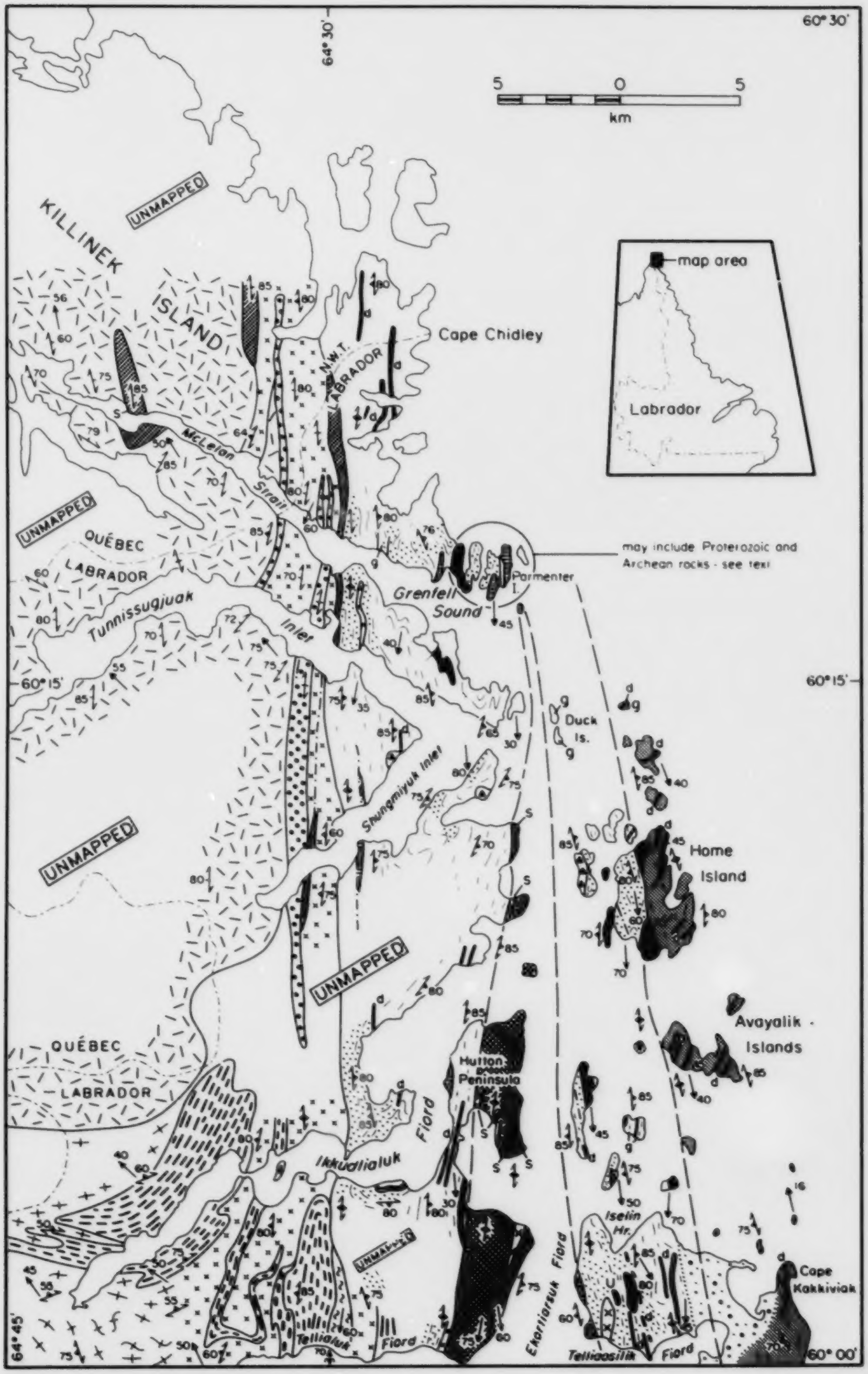
64° 30'
60° 30'
5
0
5
km
map area
Labrador
Cape Chidley
KILLINEK
ISLAND
UNMAPPED
N.W.T.
LABRADOR
McLelan
Strait
QUÉBEC
LABRADOR
Tunnissugjuak
Inlet
Grenfell
Sound
Parmenter
I.
may include Proterozoic and Archean rocks - see text
60° 15'
60°15'
Duck
Is.
Shungmiyuk Inlet
Home
Island
Avayalik
Islands
Hutton
Peninsula
Fiord
Ikkudlialuk
Iselin
Hr.
Ekortiarsuk Fiord
Telliaosilik
Fiord
Tellialuk
Fiord
Cape
Kakkiviak
64° 45'
60° 00'

Figure 2. *Geology of the field area. Areas shown as unmapped will be completed in the 1992 field season.*

LEGEND (for Figure 2)

⁘ Fluvioglacial drift

LOWER PROTEROZOIC

BURWELL TERRANE METAIGNEOUS ROCKS

⁙ Pink megacrystic granite: *foliated to mylonitic*

▨ Buff hypersthene–granodiorite: *weakly foliated to gneissic*

⁚ Grey diorite–quartz-diorite–tonalite (DQT suite): *moderately to strongly foliated and gneissic*

▨ Grey to buff hypersthene-bearing tonalite gneiss

BURWELL TERRANE METASEDIMENTARY ROCKS

▨ Granitic migmatite: *derived from quartzofeldspathic metasedimentary gneiss*

▨ Grey, banded, quartzofeldspathic psammitic gneiss: *including minor mafic (volcaniclastic?) and pelitic bands*

▨ Rusty-weathering garnet–biotite pelitic and psammitic gneiss, ± sillimanite and graphite: *generally strongly migmatitic. Contains minor interlayers of green quartzite and banded clinopyroxene–amphibolite (metavolcanic/volcaniclastic?)*

▨ Tasiuyak gneiss: *interbanded white garnet-granite and rusty biotite–garnet–sillimanite-graphite pelitic gneiss*

d Metadiabase to amphibolite dykes

LOWER PROTEROZOIC and/or ARCHEAN

× Grey diorite–metagabbro plutons: *locally gneissic and migmatitic*

ARCHEAN (Nain Crust)

▴▴ Metagabbro: *weakly to moderately deformed, locally migmatitic; may include younger rocks on Home Island*

g Pink granite: *moderately to strongly foliated*

NAIN GRANITOID GNEISSES:

▨ Buff tonalitic–granodioritic gneisses at granulite facies

▨ Grey to buff, tonalitic–granodioritic gneisses at granulite facies, partially retrograded to amphibolite facies: *locally straightened and mylonitic*

▨ Grey tonalitic–granodioritic gneisses at amphibolite-facies gneisses: *contain abundant pink granite, granite gneiss and pegmatite; locally straightened and mylonitic*

■ Anorthosite, anorthositic gabbro and anorthositic gneiss: *minor layered amphibolite/metagabbro and ultramafic rock: strongly foliated to gneissic*

U Ultramafic rocks: *generally altered to anthophyllite–actinolite assemblages*

■ Layered to massive amphibolite–mafic granulite: *generally migmatitic and includes rocks of probable plutonic and supracrustal origin*

▨ Rusty garnet, biotite, ± sillimanite and graphite pelitic to psammitic gneiss, interlayered with garnet-clinopyroxene-, ± hypersthene, amphibolite/mafic granulite gneiss

SYMBOLS

——	*Geological contact*	⟵	*Foliation*
— —	*Inferred or extrapolated contact*	⟶	*Lineation*
—·—	*Ultramylonite–mylonite shear zone*	⟶	*Minor fold plunge*
⟵	*Gneissic layering and foliation*	s	*Rusty sulphide zone*

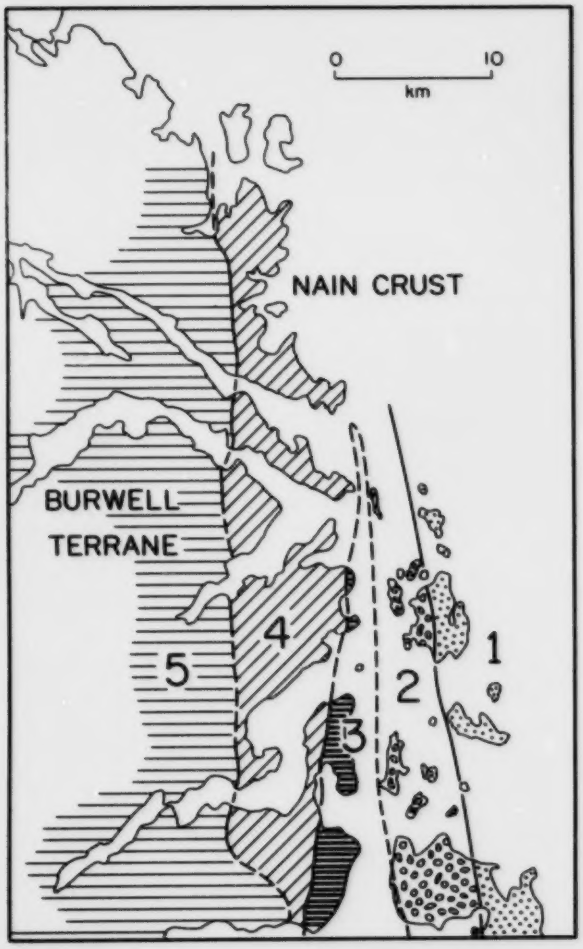

Figure 3. *Structural zonation of the area shown in Figure 2.*

ARCHEAN GNEISSES (ZONES 1, 2 AND 4)

The composition of the Archean crust does not differ significantly across the area, with the exception of the anorthosite in Zone 3. There are systematic variations in structural and metamorphic character but these are discussed in a later section.

The prevalent rock type throughout the gneiss complex is a buff- to grey-weathering, strongly migmatitic granitoid gneiss of overall tonalite–granodiorite composition (Plate 1). This appears to have been uniformly metamorphosed and migmatized in an Archean (pre-dyke) granulite-facies event and then heterogeneously retrogressed to amphibolite facies during Early Proterozoic deformation. Relict Archean gneissic texture is characterized by a profusion of disrupted enclaves of mafic and ultramafic rock enclosed within a tonalite–granodiorite leucosome (Plate 1). Strongly disrupted, so-called 'exploded' textures are common in these migmatites. Narrow screens of rusty metasedimentary gneiss are commonly associated with bands of mafic granulite and ultramafic rock. The mafic rocks comprise massive to weakly layered amphibolite or mafic granulite, probably derived from a plutonic protolith; and more irregularly layered rocks that may have originated as metavolcanic rocks. The ultramafic enclaves are commonly podiform and range in size from centimetres to kilometres in length. They consist of two general types: a yellow-weathering, heavily serpentinized rock, probably derived from peridotite; and a brown-weathering, metapyroxenite variety that is altered to anthophyllite–actinolite assemblages.

Plate 1. *Migmatitic, tonalitic, granulite-facies Archean gneiss containing disrupted mafic layers; from Zone 4. Straightening fabric at left and right of plate is due to superimposed Proterozoic strain and is associated with partial retrogression to amphibolite facies.*

The original Archean character of the gneisses is best preserved in Zone 1, where the gneisses are uniformly at granulite facies, and where hypersthene porphyroblasts are preserved in the leucosome fraction. The thorough granulite-facies migmatization is probably correlative with the ca. 2.8 Ga event defined in the Saglek Fiord area (Collerson *et al.*, 1982), which was responsible for the Late Archean metamorphism of much of the northern Nain crust. Migmatization has obscured evidence of any earlier crustal history in the Archean gneisses, although Van Kranendonk and Scott (1992) have reported possible indications of relict Early Archean crust in the area to the south.

The area around Parmenter Island (outlined on Figure 2) requires special comment. Although shown as Archean, this area contains supracrustal rocks similar to the Tasiuyak gneiss (see below) and also some mesocratic to melanocratic, locally megacrystic, tonalites that are cut by unusual green mafic dykes. Relationships between these rocks and the typical Early Proterozoic dykes have not been observed and it is possible that the Parmenter Island area includes a mixture of Archean and Early Proterozoic rocks.

ARCHEAN ANORTHOSITE (ZONE 3)

Previous work by Taylor (1977) indicated a series of anorthosite lenses along the western margin of the Archean gneiss complex. Mapping in 1991 has shown that the northern lenses are likely connected as a single pluton that underlies Zone 3 as a north-tapering wedge, and that also extends well to the south (Van Kranendonk and Scott, 1992). Although Taylor (1977, 1979) inferred an Early Proterozoic age for the anorthositic rocks, the fact that they are cut by deformed mafic dykes requires an Archean age.

The unit consists of variably foliated anorthosite, leuconorite and their gneissic equivalents, characterized by a thoroughly recrystallized, granoblastic texture containing hornblende and garnet as the principal mafic minerals (Plate 2). Relict, primary, coarse-grained texture is locally recognizable in the form of large hypersthene crystals rimmed by hornblende, and also as aggregates of granoblastic plagioclase evidently derived from the recrystallization of larger primary crystals. Thin (1 to 30 m) bands of layered amphibolite–mafic granulite, locally containing disrupted ultramafic layers, form a minor component within the unit. Another variety of anorthosite–leuconorite is a streaky, white gneiss that is injected by diffuse veins of white, garnet–hypersthene-bearing tonalite or trondhjemite. The quartz-rich composition and granulite-facies character of the tonalite–trondhjemite veins suggest that they were injected into the anorthosite gneiss during Late Archean granulite-facies melting of the host gneisses. The hypersthene migmatite fabric is cut by Early Proterozoic dykes (Plate 3) and hence of Archean age.

Plate 2. *Archean metanorthosite of Zone 3. Dark areas are aggregates of hornblende after primary orthopyroxene.*

The contacts of the anorthosite pluton are marked partly by shear zones, and partly by a 20- to 200-m-wide border sequence consisting of layered, mafic metaplutonic rocks, interlayered with anorthosite. This border sequence also includes screens of granulite-facies granitoid gneiss, mafic granulites, ultramafic rocks and minor metasedimentary gneisses, all of which are likely part of the Archean gneiss complex. The prevalence of supracrustal rocks in these border zones suggests that the anorthosite may have been preferentially intruded into such rocks, a feature that has also been noted from other Archean anorthosites (e.g., Myers, 1985).

Plate 3. *Early Proterozoic amphibolite dyke cutting granulite-facies migmatite texture in anorthositic gneiss of Zone 3.*

LATE ARCHEAN INTRUSIONS

Crosscutting units of pink to white granite and pegmatite are locally abundant and are presumed to be of Late Archean age. These crosscut granulite-facies gneissic banding but are affected by amphibolite-facies fabrics. Larger, more homogeneous, masses of Late Archean, foliated granite have been recognized on the Duck Islands (Figure 2) where they are cut by Early Proterozoic dykes. Other Late Archean intrusions are represented by small units of ophitic gabbro on Home Island and on the north shore of Shungmiyuk Inlet. These postdate most of the migmatization in the gneiss complex but have been variably deformed and locally migmatized prior to Early Proterozoic dyke injection. Gabbro on Home Island, although similar in texture to other gabbros, is not seen to be cut by dykes and could be Early Proterozoic in age.

GRANITOID PLUTONS OF UNCERTAIN AGE

The Archean gneisses are intruded by several small, lenticular plutons ranging from hornblende gabbro–diorite to tonalite in composition. These plutons crosscut gneissic fabrics in their host rocks and locally contain xenoliths of gneiss (Plate 4). They generally contain a single amphibolite-facies foliation that grades locally into a migmatitic texture. These rocks resemble plutonic rocks of the Burwell Terrane (Zone 5); however, relationships with the Early Proterozoic dykes have not been determined and their age is uncertain.

THE EARLY PROTEROZOIC DYKE SWARM (ZONES 1 TO 4)

Dykes are found throughout the Nain crust (Zones 1 to 4 inclusive), except in the very west where they die out within a distance of 2 km to 200 m of the Burwell Terrane. The dykes consist of two principal types: 1) undeformed metadiabase dykes in Zone 1 termed Avayalik-type (after Avayalik Islands), and 2) amphibolite dykes that are found throughout Zones 2, 3 and 4.

Plate 4. *Amphibolite and anorthositic gneiss inclusions in tonalitic pluton of uncertain Late Archean or Early Proterozoic age (Zone 3).*

The Avayalik-type dykes are typically dark brown but vary to black or dark green, have well-preserved chilled margins, and contain scattered phenocrysts of black plagioclase. Some of the thin dykes are affected by marginal cleavage development, but the majority are straight and undeformed (Plate 5). In detail, they possess a fine, granoblastic texture composed of hornblende and plagioclase intergrown with scattered clusters or strings of garnet and local hypersthene porphyroblasts. The dykes have a consistent northeast strike and moderate southeasterly dip that is aberrant in comparison with the east-west trends and vertical attitudes of dykes in Nain Province farther to the south (Taylor, 1979). However, equivalent areas of West Greenland (allowing for pre-drift reconstruction) contain abundant northeast-trending dykes, (Nielsen, 1987), though with vertical dips.

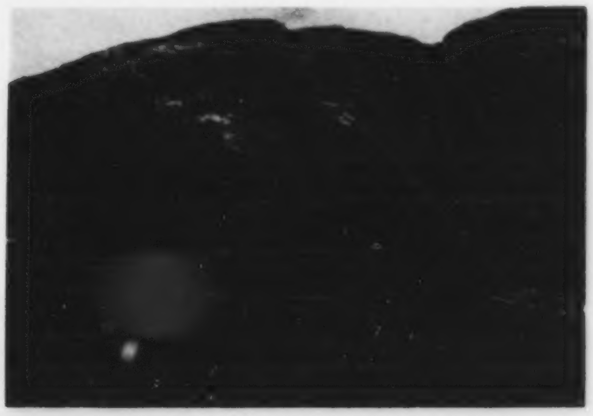

Plate 5. *Avayalik-type Early Proterozoic dykes crosscutting Archean granulite-facies gneiss in Zone 1. Notebook for scale on left-hand dyke.*

The amphibolite dykes are generally black or green and are commonly, but not invariably, porphyritic, with individual dykes containing up to 20 percent white plagioclase crystals or grain aggregates. The dykes have been pervasively recrystallized to granoblastic or foliate textures defined by hornblende-clinopyroxene ± garnet assemblages, but lack migmatitic textures. Additional varieties, seen mostly within the anorthosite unit, are garnetiferous dykes, distinguished by large garnet–plagioclase ± hornblende symplectites, and dykes containing large, rounded plagioclase phenocrysts up to 3 cm across. North-south trends predominate in all dykes and are the result of straightening, folding and shearing during Proterozoic deformation (Plate 6). Transitional stages have been observed between the porphyritic amphibolite dykes and Avayalik-type dykes, but the equivalence of the two types will have to be further tested through geochemical studies. Although all mafic dykes are assumed to be Early Proterozoic, there may have been more than one age of injection, as is the case in the Nachvak Fiord area to the south, where two sets of dykes have been recognized (Wardle, 1983).

Plate 6. *Early Proterozoic amphibolite dyke cutting retrogressed, granulite-facies Archean gneiss has been partially transposed into the north-south structural trend of Zone 2.*

BURWELL TERRANE, GRANITOID ROCKS (ZONE 5)

The Burwell Terrane is dominated by granitoid rocks and their gneissic derivatives. These are grouped into two principal suites: 1) a large, relatively homogeneous hypersthene–granodiorite batholith that occupies most of the northern part of Zone 5, and extends well to the west, and 2) a more varied suite of dioritic, quartz dioritic and tonalitic rocks of broadly calc-alkaline character, referred to as the DQT suite, that occurs in the east and south.

Hypersthene–Granodiorite

This is typically a buff- to brown-weathering rock of predominantly granodioritic composition but varying locally to tonalite. A K-feldspar megacrystic phase is also present in a small area around central Tunnissugjuak Fiord. Where retrogressed within shear zones, this rock type looks remarkably similar to the megacrystic granite sheets that intrude the DQT suite (see below). The hypersthene–granodiorite varies from strongly foliated to gneissic, but in low strain areas, primary features such as melanocratic (cognate?) xenoliths in various stages of assimilation (Plate 7), and convolute syn-plutonic mafic dykes are preserved. Partially assimilated, slab-like inclusions of mafic granulite gneiss are also abundant and contain a fabric that at least locally can be demonstrated to predate their incorporation in the granodiorite. The eastern contact of the granodiorite with the DQT suite is mostly occupied by younger, megacrystic granite sheets. Its southern contact has not been mapped but is extrapolated (Figure 2) on the basis of the unit's distinctive aeromagnetic expression.

Plate 7. *Hypersthene–granodiorite of Burwell Terrane (Zone 5) showing relict igneous texture and also cognate xenoliths in various stages of assimilation.*

The DQT Suite

The easternmost part of this suite is seen as irregular dykes of dark diorite to tonalite that are intruded across migmatite fabrics in the Archean gneisses (Plate 8). The dykes do not extend more than 200 m into the Nain crust; however, the plutons of uncertain age described above may represent more easterly examples of the suite. The DQT suite immediately west of the Nain crust consists of strongly foliated to banded, grey, dioritic–tonalitic gneisses. These grade westward into more massive hypersthene-bearing, augen gneiss variants that retain relict plagioclase-porphyritic texture. Orthopyroxene-bearing varieties of the DQT suite predominate north of Tunnissugjuak Inlet, whereas hornblende-bearing varieties are more common to the south.

Plate 8. *Metatonalite dyke of DQT suite cutting Archean migmatite gneiss fabric; western boundary of Zone 4.*

Relict plutonic textures are best preserved in more southerly parts of the DQT suite, notably between Shungmiyuk Inlet and Tellialuk Fiord. Multiple intrusive phases are present consisting of early enclaves of dark, plagioclase-porphyritic hornblende gabbro or diorite cut by many different phases of quartz diorite and tonalite. The youngest phases are generally leucotonalite and pegmatite dykes (Plate 9). The DQT suite plutonic rocks, which generally contain only a single foliation, pass gradationally into zones of banded, migmatitic gneiss that locally preserve orthopyroxene porphyroblasts in their leucosome. Similar rocks have been described by Van Kranendonk and Scott (1992) from the area to the south.

In the northern part of the area, the DQT suite is intruded by numerous thin sheets of pink, megacrystic granite and derived augen gneiss, of which only the larger examples are shown on Figure 2. These have localized much of the Proterozoic strain and are generally strongly sheared, and locally converted to a distinctive pink and black ultramylonite containing relict K-feldspar porphyroclasts.

Plate 9. *Multiple intrusive phases in DQT suite, Burwell Terrane (Zone 5). Dark inclusions of diorite are set in a polyphase matrix of paler quartz diorite–tonalite and are intruded by leucotonalite and pegmatite dykes.*

Southwest of Ikkudlialuk Fiord, the recognizable plutonic rocks of the DQT suite pass westward along strike into strongly foliated, hypersthene-bearing equivalents, and then into a buff, well-banded, granulite-facies gneiss of overall tonalitic composition. The transition is thought to be solely the result of progressive metamorphism and deformation, but further field work is required before this can be confirmed.

The plutonic rocks of Burwell Terrane have yet to be dated. The westernmost parts of the DQT suite clearly postdate migmatization in the Archean gneisses but relationships have not been observed with the Proterozoic dyke swarm, which does not extend into the area intruded by DQT rocks. However, by comparison with the observations of Van Kranendonk and Scott (1992), who have described analogues of the DQT suite crosscutting Early Proterozoic mafic dykes, an Early Proterozoic age for the DQT suite is inferred. As well, indications of a Proterozoic age are provided by preliminary Nd model (depleted mantle) ages of 2.9–2.1 Ga (Jackson and Hegner, 1991), which have been interpreted as mixing of juvenile Proterozoic igneous material and remelted Archean crust. The samples analyzed by Jackson and Hegener (1991) came from the archived collection of Taylor (1979), but their field locations are unknown.

BURWELL TERRANE METASEDIMENTARY GNEISSES (ZONE 5)

Zone 5 contains a variety of metasedimentary gneisses, all of presumed Early Proterozoic age, and distinct from the Archean metasedimentary gneisses lying to the east.

Metasedimentary gneiss north of Shungmiyuk Inlet occurs as a discontinuous strip along, or near, the contact with the Archean gneisses, and as a folded belt within the hypersthene–granodiorite of Killinek Island. The gneiss is a well-layered rock consisting of alternating white, garnetiferous granite and rusty, biotite–garnet–graphite ± sillimanite pelitic gneiss, correlated with the Tasiuyak gneiss described from areas to the south (Wardle, 1983; Goulet and Ciesielski, 1990). The contact between the Archean gneiss complex and Tasiuyak gneiss is straightened but not mylonitized.

Metasedimentary gneisses south of Shungmiyuk Inlet are of rather different aspect. The most prominent are extensive belts of rusty-weathering, pelitic–semipelitic gneiss containing biotite–garnet–graphite ± sillimanite assemblages and found around Ikkudlialuk Fiord. These are generally strongly migmatitic with 20 to 50 percent granite and pegmatitic veins, and contain layers of garnetiferous amphibolite and distinctive green, garnetiferous quartzite. Some outcrops also contain thin bands of a plagioclase-porphyritic amphibolite (Plate 10) that locally preserves relict grading and may have originated as volcanic layers in the gneiss protolith.

Plate 10. *Well-layered supracrustal gneiss composed of alternating dark amphibolite bands (locally plagioclase-porphyritic) and pale quartzofeldspathic gneiss from Burwell Terrane (Zone 5), north shore of Ikkudlialuk Fiord.*

A metasedimentary rock type exposed on the north shore of Tellialuk Fiord consists of grey, well-banded, quartzofeldspathic gneiss and its migmatitic equivalents (Plate 11). Plagioclase-porphyritic amphibolite layers, resembling those described above, are also present within this unit. Mapping in the area between Ikkudlialuk and Tellialuk fiords

Plate 11. *Layered quartzofeldspathic gneiss of psammitic origin from Burwell Terrane (Zone 5), south shore of Ikkudlialuk Fiord. Hammer is 50 cm long.*

is incomplete, and the distribution of metasedimentary gneisses shown on Figure 2 is speculative.

POST-OROGENIC ROCKS

Several lamprophyric dykes were found at the western end of Tunnissugjuak Inlet, and around the entrance to Ikkudlialuk Fiord. Scattered northeast-trending diabase dykes occur west of Ikkudlialuk Fiord. All of these dykes are posttectonic but their age is otherwise unknown.

TECTONIC EVOLUTION

The structure of the area is described from east to west in terms of the lithotectonic zones established in Figure 3.

Zone 1

The crosscutting nature of the dykes (Plate 5) in this zone indicates that virtually all major fabric development is of Archean age. The earliest fabric is an irregular gneissic layering that is disrupted and injected by a granulite-facies leucosome, which commonly contains orthopyroxene porphyroblasts partially rimmed by hornblende. Gneissic fabrics are affected by a set of southeast-plunging, tight to isoclinal folds, associated with the development of a weak axial-planar biotite–orthopyroxene fabric and an orthopyroxene lineation, indicating that this event also occurred at granulite facies (Figure 4a). The second set of folds appears geometrically related to north-south-trending mylonitic shear zones, defined by amphibolite-facies assemblages, that are crosscut by the Early Proterozoic dykes (Figure 4a). The only Proterozoic fabrics identified within Zone 1 are weak, hornblende fabrics on dyke margins and thin (< 20 cm) ultramylonite zones with 340° trends.

Zone 2

The transition from Zone 1 to 2 takes place over less than 100 m and is defined by the appearance of retrogressed, straightened, grey gneisses and mylonitic, ductile shear zones that affect both gneisses and dykes (Figure 4b; Plate 6). Most of

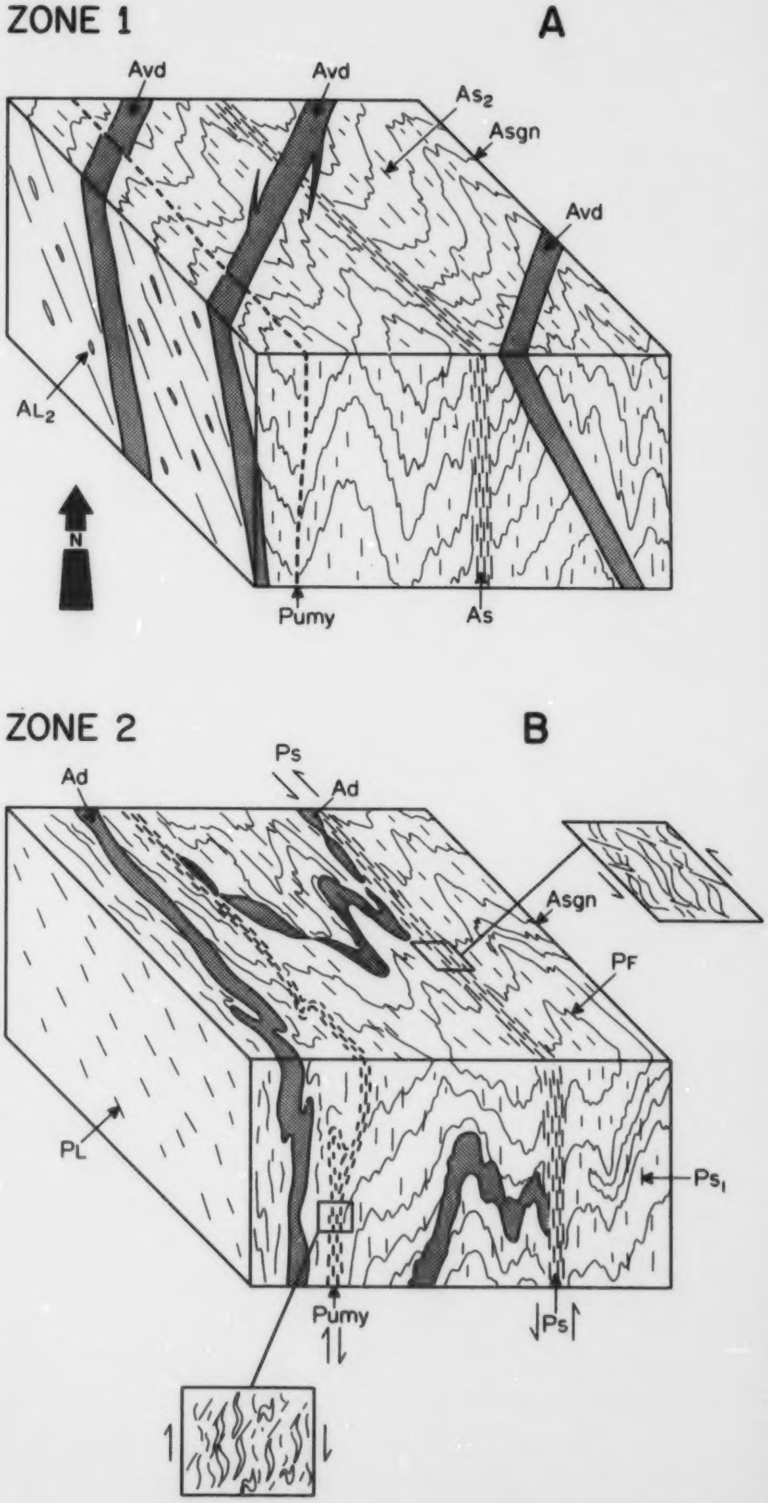

Figure 4. *Schematic summaries of structural geometry and deformation kinematics in structural Zones 1 (A) and 2 (B) (see Figure 3). Insets show shear sense of kinematic indicators. ASgn–Archean gneissosity; As–Archean shear zone; Avd–Avyalik-type dyke; AS$_2$–Archean second fabrics; AL$_2$–Archean (second) linear fabric; Ad–amphibolite dyke; Pumy–Proterozoic ultramylonite zone; Ps–Proterozoic shear zone; PF–Proterozoic fold; PS$_1$–Proterozoic axial-planar fabric; PL–Proterozoic linear fabric.*

Zone 2 is dominated by Early Proterozoic straightening and ductile shearing associated with folding and reorientation of the Proterozoic dyke swarm into north-south trends. Proterozoic folding of Archean gneissosity is widespread. The folds generally show S-sense asymmetry and typically plunge steeply to the south (Figure 4b), and more locally to the north. The Proterozoic structures can be seen locally to refold earlier folds that are presumably Archean F2 structures (Figure 4b). The axial-planar fabric to the Proterozoic folds, which is typically defined by biotite and/or hornblende alignment, trends 350° and is subvertical. Proterozoic mylonite zones, developed synchronously with folding, are characterized by fine-grained, grey, mylonitic gneiss that passes into thin zones (< 3 m) of black, glassy ultramylonite containing porphyroclasts of plagioclase and locally hornblende and garnet. A strong mineral lineation associated with the mylonite zones plunges steeply south to southeast, coaxial with Proterozoic fold hinges. The shear zones range up to 100 m in width and probably form a braided system isolating islands of less-deformed rock. This is best seen in the area south of Iselin Harbour (Figure 2) where a large, strain augen of Archean granulite with S-sense asymmetry is bounded by mylonitic shear zones.

Kinematic indicators suggest a complex movement history. Shear bands and rotated porphyroblasts seen in plan view (normal to lineation) invariably reveal a sinistral sense of motion (Plate 12), locally coupled with evidence in vertical section for a component of east-side-up motion. This is similar to the observations of Van Kranendonk and Scott (1992) in the area to the south. Rotated feldspars and shear bands seen in vertical sections of the glassy ultramylonite zones, parallel to lineation, usually demonstrate a west-side-up sense of motion (Plate 13). The ultramylonites locally crosscut other structures and are provisionally considered to represent a later period of shearing developed at higher crustal levels.

The mylonite zones and intervening gneisses have been locally affected by a late set of Proterozoic north-south-trending folds characterized by gentle plunges and lack of penetrative axial-planar fabrics.

Zone 3

The anorthosite of this zone locally preserves evidence of an Archean (pre-dyke) granulite-facies migmatization (Plate 3), but has been pervasively refoliated by amphibolite-facies Proterozoic deformation. Proterozoic fabrics trend north-south and are associated with south-plunging folds and mineral lineations, which are generally collinear with those in Zone 2. A pervasive sense of sinistral shear is revealed by folds having a consistent S-sense of asymmetry and by rotated garnet–hornblende–plagioclase symplectic aggregates (Plate 14). Development of mylonite is a minor feature that is limited to thin zones at the margins of the anorthosite, and thin bands (< 1 m) of black, glassy ultramylonite that form local conjugate sets trending 065° and 120°. The discordant nature of the ultramylonites is unique to Zone 3 and illustrates the late nature of this style of deformation, and the relative competency of the anorthosite.

Plate 12. *Sinistral shear bands seen in mylonitic gneiss; Zone 2. View is of horizontal surface; lineation plunges steeply to south.*

Plate 13. *Ultramylonite seen in vertical section (parallel to lineation), looking north in Zone 2. Fine shear bands dip to left of plate and indicate west-side-up displacement.*

Zone 4

The structure of this broad zone is characterized by a large component of well-preserved, Archean granulite-facies gneiss, which has been locally retrogressed in zones of

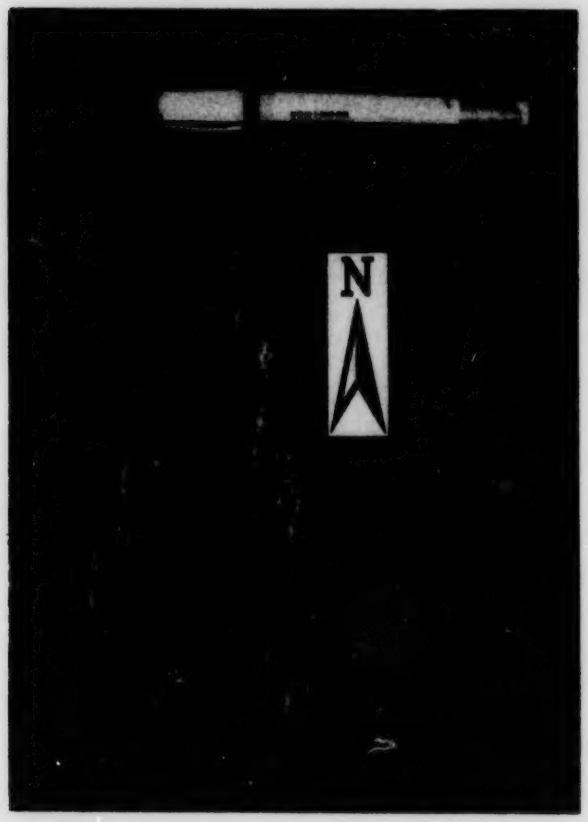

Plate 14. *Sinistral rotational textures in garnet–plagioclase–hornblende symplectites, Zone 4. View is of horizontal surface; lineation plunges steeply to south.*

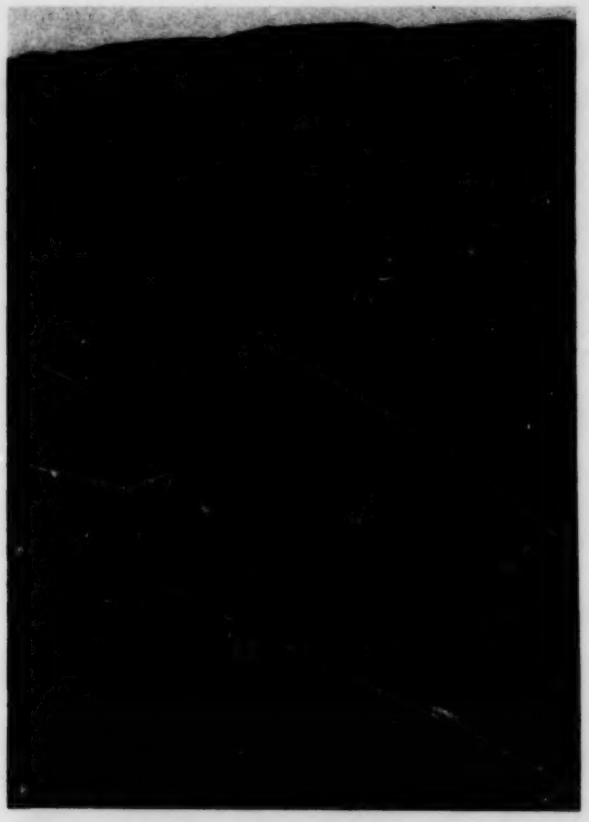

Plate 15. *Straightened, granulite-facies Archean gneisses, partially retrogressed to amphibolite facies, from Tunnissugjuak Inlet, western end of Zone 4. Height of cliff is about 100 m.*

Proterozoic straightening (Plate 1). Archean gneissic fabrics are folded into a series of major, south-plunging folds associated with a weak to strong mineral lineation collinear with those in zones to the east. Ductile shearing and ultramylonite development is only strongly developed on the western boundary of the zone (Plate 15). Amphibolite-facies retrogression in the Archean gneisses also becomes pronounced in the west, within 500 m of Zone 5.

Amphibolites in Zones 2, 3 and 4, including layers in the Archean gneisses, and the Proterozoic dykes, have developed garnet–plagioclase–hornblende symplectites. These may reflect decompressional effects during Proterozoic metamorphism.

Zone 5

The boundary of Zones 4 and 5 is generally, but not invariably, marked by thin, discontinuous shear zones. North of Shungmiyuk Inlet these are preferentially located within screens of megacrystic granite, where they are seen as pink and black, feldspar-porphyroclastic ultramylonite zones texturally similar to those of Zone 2. The Archean Tasiuyak gneiss contact is straightened but only weakly affected by mylonitization. This contrasts with areas to the south where it is the locus of the major Abloviak shear zone (Wardle, 1983). Mylonitization south of Shungmiyuk Inlet is restricted to thin (< 50 m) zones localized along the Zone 5–Zone 4 boundary. Lineation in these mylonite zones is poorly developed but, where present, has a steep down-dip plunge. Sparse kinematic indicators suggest a west-side-up sense of movement.

The narrow mylonite zones give way westward to a 4-km-wide, north-south-trending belt of straightened gneiss that dips steeply to the west. These straightened gneisses die in the interior of Zone 5, which is dominated by irregular, folded-fabric trends that appear, on the basis of incomplete mapping, to be related to a set of late, northwest-plunging folds associated with a moderate northwest-plunging orthopyroxene-hornblende lineation. To the south, in the vicinity of Ikkudlialuk Fiord, there is a major swing in strike to east-west trends, accompanied by the development of complex interference fold patterns. As the east-west trends are traced toward the coast they swing into alignment with the regional north-south trends. This deflection is interpreted to be the result of the sinistral deformation seen east of Zone 5.

Ubiquitous, stable hypersthene attests to the attainment of Proterozoic granulite-facies conditions in the interior of Zone 5. Amphibolite-facies assemblages prevail throughout the eastern part of Zone 5; however, the presence of relict hypersthene in rocks of the DQT suite indicates that the amphibolite-facies assemblages may be of retrograde origin, possibly associated with ductile shearing. Hypersthene-bearing assemblages are also locally retrograded to amphibolite facies along thin shear zones in the interior of the hypersthene−granodiorite batholith.

MINERALIZATION

Several pyritiferous gossan zones have been discovered within and along the margins of the Archean anorthosite of Zone 3. The mineralized zones vary from centimetres to several metres in width and follow compositional layering in both anorthosite (Plate 16) and layered metagabbro−ultramafic rocks. Their strike length locally attains several hundred metres but in most cases has not been established beyond the coastal strip. Six selected grab samples analyzed for PGE, Au, Ni and Cr yielded disappointing results, although it should be noted that the deep gossan weathering of the sulphide zones made it difficult to obtain fresh samples. A more thorough sampling program will be undertaken in 1992.

Plate 16. *Rusty, pyritiferous zone (dark layer on which notebook rests) in metanorthosite of Zone 3.*

A large area of gossan was also discovered on the north shore of McLelan Strait (NWT) along the boundary between hypersthene−granodiorite and Tasiuyak gneiss. Analytical work is in progress and will be reported at a later date (M. Van Kranendonk, personal communication, 1991). Small gossan zones, generally less than a few metres in width, were also discovered within the hypersthene−granodiorite of Burwell Terrane; however, analysis of grab samples for Au and related elements yielded negative results.

D. Wilton (personal communication, 1991) has also reported concentrations of Cr in an ultramafic component of the anorthosite suite south of Ekortiarsuk Fiord.

DISCUSSION

The results of the 1991 field work require that considerable modification be made to previous ideas regarding the nature of the Burwell Terrane and its boundary with the Nain Province. The most significant is the finding that the Burwell Terrane is composed predominantly of plutonic rocks. These are presumed to be of Early Proterozoic age and, according to the Nd work of Jackson and Hegner (1991), must be at least in part of juvenile origin. The calc-alkaline (diorite−quartz diorite−tonalite−granodiorite−granite) composition of these rocks suggests an affinity with subduction-related plutonism and an origin in the deep levels of a magmatic arc. This arc magmatism affected only the western part of the Nain crust within the map area; however, to the south Van Kranendonk and Scott (1992) have reported intrusive activity spread across the breadth of the Nain crust. For at least the later part of its history, therefore, the Early Proterozoic Burwell 'arc' must have been founded upon a substrate of Nain crust. This requires that the idea of the Komaktorvik shear zone as a fundamental suture separating the Nain Province from the Burwell Terrane must be questioned. The shear zone was originally postulated on the basis of strong aeromagnetic and airphoto lineaments but it now appears that these result largely from regional fabric straightening, rather than mylonitization. Intense ductile shearing is therefore not as important as previously suspected and clearly postdates Burwell intrusive activity. The possibility remains, however, that early components of Burwell Terrane may represent tectonically accreted crust. In order to resolve this issue it will be necessary to determine the extent, if any, of Nain crust under Burwell Terrane; and to establish the chronology of Burwell plutonism, in particular the timing of its links with the Nain crust. Delineation of the western extent of Nain crust will depend upon future Nd studies. A U−Pb dating program, presently being undertaken by D. Scott, will establish the age range of Burwell plutonism.

Apart from the Nain−Burwell boundary, the major shear system within the area is that represented by Zone 2, well within the Archean crust. In regional terms, both shear zones can probably be considered part of the Komaktorvik shear zone; indeed they may merge in areas to the south. In a provisional kinematic interpretation, it is suggested that Early Proterozoic deformation was a two-stage event. Initial folding, shearing and lineation development, concentrated within Zone

2 but also affecting Zones 3 and 4, took place in an oblique, collisional environment by sinistral shear associated with east-side-up vertical movement. This probably occurred in response to northward motion of the Nain Province, with respect to the Rae Province, following collision of the two cratons along the Abloviak shear zone (Van Kranendonk and Ermanovics, 1990; Van Kranendonk and Scott, 1992). Sinistral shear was followed by a phase of west-side-up motion and production of glassy ultramylonites in response to late cross-orogen shortening. These late ultramylonite zones generally appear to have reactivated pre-existing sinistral structures, except within the Zone 3 anorthosite, where a conjugate set of discordant ultramylonites is developed. The formation of glassy ultramylonites in response to late shortening is thought to be comparable to more southerly parts of Torngat Orogen, where similar rocks along the eastern margin of the Abloviak shear zone represent the latest events dated between 1805 and 1782 Ma (Bertrand *et al.*, 1990).

Early Proterozoic metamorphism within Zone 1 appears to have been minimal, possibly apart from the garnet−hornblende ± hypersthene assemblages developed in the Avayalik-type dykes. These require some discussion, particularly since identical dykes have been described by Morgan and Taylor (1972) from an area of Nain crust, 60 km to the south, suggesting that Avayalik-type dykes are a regional phenomenon. The assemblage may result simply from a regional, granulite-facies metamorphism; however, if this is the case, it is remarkable that the dykes show none of the boudinage, marginal shearing or pervasive foliation that would be expected in an environment of high-grade regional metamorphism and deformation. An alternative explanation is that the dykes were intruded into deep, hot crust and developed their metamorphic assemblages through a process of slow, static cooling at elevated temperatures; a process of auto-recrystallization. In this respect, there appears to be an analogy with the Early Proterozoic Kangamiut dykes of west Greenland (Windley, 1970; Bridgwater *et al.*, 1976), which have developed garnet−hornblende assemblages through a process of static recrystallization in areas of Archean granulite crust unaffected by subsequent tectonism. Although the Avayalik and Kangamiut dykes differ significantly in many aspects, and a correlation is not suggested, the process of auto-recrystallization may have been similar. The implications of both explanations, regional metamorphism and auto-recrystallization, are the same; Zone 1 must have been deeply buried and hot at the time of dyke emplacement. This is consistent with the model suggested by Van Kranendonk and Scott (1992), in which the northern parts of the Nain crust are proposed to represent deep structural levels that have been exhumed by ramping and progressive east-side-up uplift along the Komaktorvik shear system.

Early Proterozoic metamorphism in the remainder of the area appears to have increased from amphibolite facies in the east (Zone 2) to granulite facies in the interior of the Burwell Terrane. Exposure of granulite-facies rocks in the Burwell Terrane may have been aided by exhumation along the west-dipping shear zones located at the eastern edge of the terrane.

Another major revelation of 1991 field work is the unexpectedly large size of the anorthosite pluton represented by Zone 3. This is now seen as a single pluton at least 50 km long (including areas described by Van Kranendonk and Scott, 1992, to the south) whose southern limit is as yet undefined. A chain of similar anorthosite units extends a farther 120 km to the south (Taylor, 1979) indicating that these rocks are a fundamental feature of the western Nain crust. Archean anorthosites have generally been strongly dismembered by subsequent tectonism and granitic intrusions. An example is the well known Fiskenaesset complex of West Greenland (Bridgwater *et al.*, 1976; Myers, 1985) which, despite its large original size, is now preserved as a multitude of small segments, rarely more than 10 km in length. This feature is also true of most of the Nain Province anorthosites, which occur as small podiform masses rarely larger than a few km in diameter (e.g., Wiener, 1981). The size and integrity of the Zone 3 anorthosite is therefore surprising. Discussion of the origin of these anorthosites is beyond the scope of this paper, however, the idea that they represent the vestiges of early oceanic terranes appears to be the favoured model (e.g., Myers, 1981).

In the context of the map area, the most important problem posed by the anorthosite is its restriction to the western margin of the Nain crust. It is possible that this location is a fortuitous result of Archean tectonism, but it is tempting to compare it with the Kapuskasing structure of the Superior Province (Percival *et al.*, 1989) where Proterozoic thrusting has exposed an anorthosite-rich layer in the lower levels of the Archean crust. The anorthosite of Zone 3, together with the gneisses of Zone 4, may similarly have been dragged up from depth by ductile shearing within the Zone 2 shear system and along the Nain−Burwell boundary. If this model is correct there should be a systematic increase in both Archean and Proterozoic paleo-pressure conditions from Zone 2 into Zone 4. The field evidence does not offer any convincing support in this direction, since both Archean and Proterozoic assemblages are similar in all of these zones. Thermobarometric work in progress by F. Mengel may, however, shed additional light on the problem.

Finally the discoveries of pyrite-rich mineralization within both the anorthosite, and the interior of Burwell Terrane, have revealed an unexpected mineral potential, particulary for Ni and PGE, that requires follow-up. The greater than expected extent of anorthosite, coupled with the recognition that it belongs to a single pluton, is believed to enhance its potential as a new exploration target in northern Labrador.

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PRELIMINARY REPORT ON EXAMINATION OF SULPHIDE-RICH ZONES WITHIN THE PROTEROZOIC SNYDER GROUP, LABRADOR, AND THE SEDEX POTENTIAL OF THESE UNITS[1]

D.H.C. Wilton and D. Phillips
Department of Earth Sciences, Memorial University of Newfoundland
St. John's, Newfoundland A1B 3X5

ABSTRACT

The Lower Proterozoic Snyder Group is an 11- by 3-km sequence of metasedimentary rocks within the contact aureole of the Kiglapait Intrusion, northern Labrador. The group comprises five formations viz.; Lower Quartzite, Iron Silicate, Quartzite–Marble, Graphite–Sulphide Siltstone and Upper Quartzite. The Graphite–Sulphide Siltstone contains massive sulphide lenses and interlayers, which are composed predominantly of pyrrhotite and have lesser pyrite and minor chalcopyrite. The presence of sphalerite has been reported by others. The Graphite–Sulphide Siltstone formation was sampled to assess base-metal contents. The XRF analyses indicate maximum Cu, Pb and Zn contents of 1352, 237 and 3589 ppm, respectively. Pyrrhotite and pyrite are also common in a granodiorite that intruded the Snyder Group. The potential for significant concentrations of base metals within the Snyder Group appears to be low.

INTRODUCTION

This project was initiated to examine sulphide-rich sedimentary zones within the Snyder Group, Labrador. The two objectives of this part of the project are (1) to define the economic potential of such zones, and (2) compare them with known sedimentary exhalative (SEDEX) deposits such as the Sullivan Pb–Zn deposit of British Columbia (Hamilton *et al.*, 1983). The study is part of a larger project to examine the SEDEX potential of Proterozoic sedimentary sequences developed on the Archean Nain Province of northern Labrador. The other units of interest include the Mugford and Ramah groups, which will be examined during the 1992 and 1993 field seasons, respectively.

The Snyder Group constitutes a Proterozoic, predominantly sedimentary, supracrustal sequence that unconformably overlies Archean basement gneiss to the northwest of the Middle Proterozoic (ca. 1.32 Ga) Kiglapait Intrusion (Figure 1), and is essentially within the contact-metamorphic aureole of the intrusion (Speer, 1978). The Snyder Group is exposed south from Cold Comfort Bay in a very limited outcrop pattern that is only about 11 km long and 2 to 3 km wide. Small, isolated remnants of the unit extend along the southern coast of Snyder Bay for another 15 km east from Cold Comfort Bay (Speer, 1978).

PREVIOUS WORK

The sedimentary components of this unit were first described by Wheeler (1942), who concluded they were younger than the surrounding gneisses. Morse (1969) formally defined the rocks as the Snyder Group. Speer (1972), mapping predominantly in the Cold Comfort Bay area, designated a stratigraphy for the unit consisting of five members *viz.;* (1) Lower Quartzite member, which unconformably overlies the Archean gneiss with a basal conglomerate, (2) Quartz-Gneiss member, up to 10 m thick, which consists of alternating quartz, quartz–amphibolite–garnet, quartz–cordierite and quartz–hypersthene layers, (3) Quartz–Marble member, (4) Sulphide Hornfels member, and (5) Upper Quartzite member. Speer (1972) defined the area between the Snyder Group and the Kiglapait Intrusion as the Outer Border Zone of the Kiglapait Intrusion, and also mapped the Wendy granodiorite (so named by Morse, 1969) as intruding along the margin between the Snyder Group and the Outer Border Zone south of Cold Comfort Cove.

Berg (1975) remapped the Snyder Group in the Cold Comfort and Middle bays area and completed more detailed mapping of the unit south to the Avakutakh River. He renamed the middle members of the Snyder Group, the Silicate–Iron Formation, the Quartzite–Marble and the Graphite–Sulphide Hornfels, and he suggested a redefinition of the Snyder Group, in which the Snyder Group as mapped by Speer (1972) was designated as the Lower Snyder Group and the Outer Border Zone became the Upper Snyder Group. The reason for this subdivision into upper and lower groups was that Berg *(op. cit.)* could recognize distinct, mappable units, in particular a banded-ironstone unit and a calcsilicate unit, stratigraphically above Speer's Upper Quartzite member. The

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[1] Project funded as grant to D.H.C. Wilton from the Labrador Comprehensive Agreement (ACOA and Enterprise Newfoundland and Labrador)

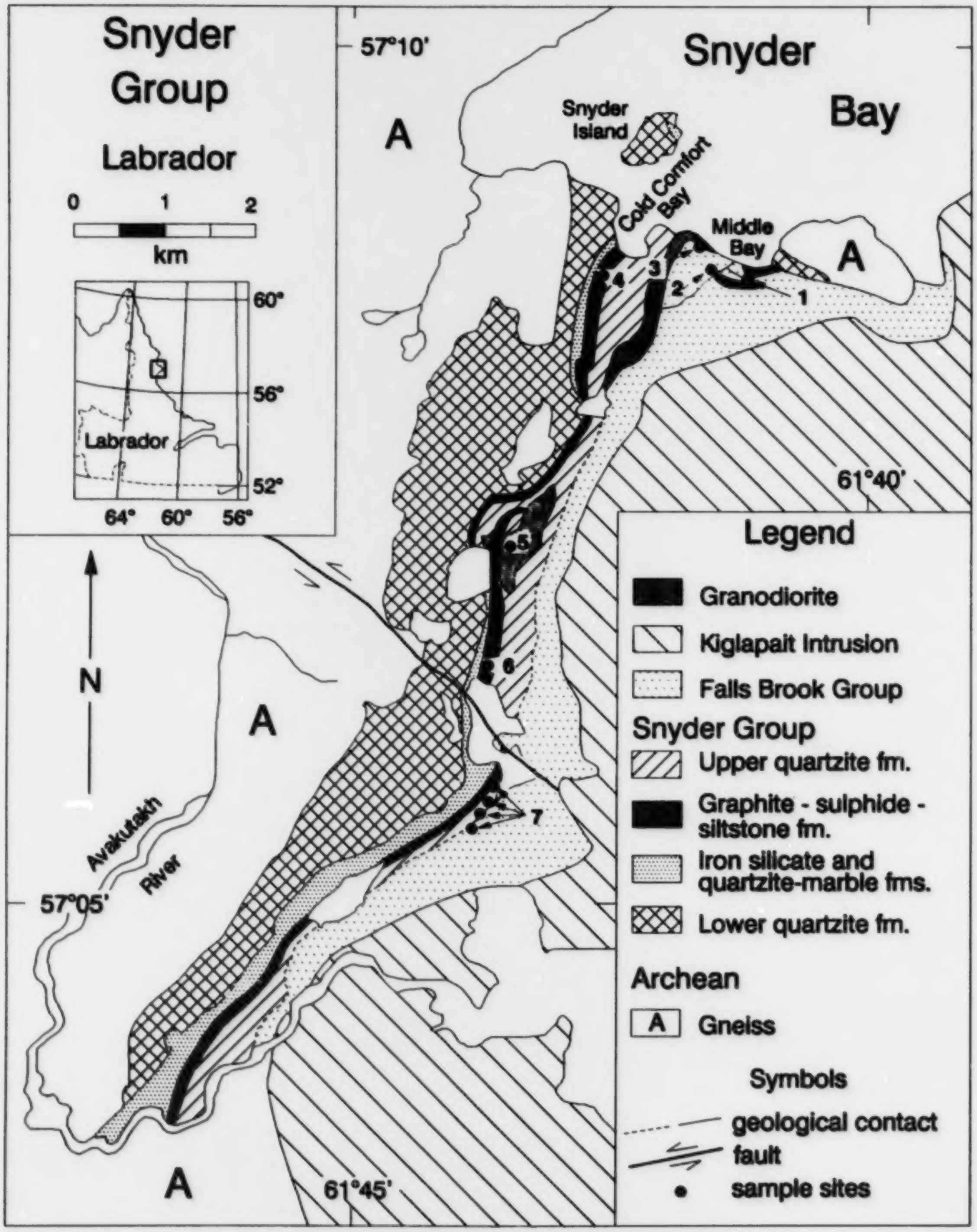
Snyder Group
Labrador
0
1
2
km
60°
56°
52°
Labrador
64°
60°
56°
57°10'
Snyder Bay
Snyder Island
A
Cold Comfort Bay
Middle Bay
1
2
3
4
5
6
7
61°40'
N
A
Avakutakh River
57°05'
A
61°45'
Legend
Granodiorite
Kiglapait Intrusion
Falls Brook Group
Snyder Group
Upper quartzite fm.
Graphite - sulphide - siltstone fm.
Iron silicate and quartzite-marble fms.
Lower quartzite fm.
Archean
A Gneiss
Symbols
geological contact
fault
sample sites

Figure 1. *Geology of the Snyder and Falls Brook groups, Labrador; compiled from Berg (1975) and Speer (1978). Sample-site localities are denoted as solid circles.*

type area for the Upper Snyder Group is along Falls Brook, which flows into Middle Bay. He was unable to define the contact between Upper and Lower Snyder groups (although he suggested that it might be unconformable) because of the presence of extensive mafic granulites within the Upper Snyder Group, which obscured the contact zone and expanded the stratigraphic width of the upper unit. He was also unable to readily distinguish if the granulites represented metavolcanic rocks or were sills. Berg's *(op. cit.)* mapping also revealed that the Graphite–Sulphide Hornfels and Upper Quartzite members extended south to the Avakutakh River.

Speer (1978) modified his stratigraphy of the Snyder Group and suggested that the stratigraphic thickness of the unit was on the order of 240 m. He described the constituent formations as A formation: lower quartzite (120 m thick), B formation: iron formation (15 m), C formation: quartzite–marble (10 m), D formation: graphite–sulphide siltstone (14 m), and E formation: upper quartzite (80 m). Speer's nomenclature is used in this report except for the alphabetical annotation for the formations (e.g., the lowest unit is referred to here as 'Lower Quartzite' rather than 'A formation'; the second formation is called the 'Iron Silicate formation' rather than 'Iron formation').

Speer (1978) continued to refer to the rocks between the Snyder Group and the Kiglapait Intrusion as the Outer Border Zone, but in discussing Berg's findings, he suggested that the sedimentary units within the zone must be unconformable upon the Snyder Group. He also renamed the Wendy granodiorite between the Snyder Group and the Outer Border Zone as the Border breccias, and postulated, as had Morse (1969), that they represent anatectic melts of country rock caused by the Kiglapait Intrusion. Speer *(op. cit.)* found that the granodiorite resembles the upper quartzite formation and the two units are commonly difficult to distinguish between in the field.

Smyth and Knight (1978) compared the stratigraphic columns of the Snyder, Mugford and Ramah groups. They suggested that the Snyder Group was correlative with the other two groups up to and including the Graphite–Sulphide Siltstone formation, but that the Upper Quartzite formation was unique. They generally agreed with Speer's (1978) five-fold subdivision of the Snyder Group, but suggested that describing the B formation as an iron formation is a misnomer as the unit actually contains less than 15 percent Fe, and hence is not iron formation *sensu stricto*.

Docka (1980), mapping the mafic granulites in Speer's (1978) Outer Border Zone or Berg's (1975) Upper Snyder Group, called the host unit the unnamed sequence. Docka *(op. cit.)* was unable to define an exact protolith for the granulites but suggested that they predated the Kiglapait Intrusion. He agreed with Berg's (1975) revision of the Snyder Group south of Cold Comfort Bay in that Graphite–Sulphide Siltstone and Upper Quartzite formations crop out as far as the Avakutakh River. Also, he further suggested that the granodiorite along the eastern side of Cold Comfort Bay (Morse's [1969] Wendy granodiorite) is an igneous breccia, which may have been intruded along the fault that juxtaposed the Upper Quartzite formation with southeast-trending units of his unnamed sequence.

Schuh (1980a), in examining the geology of the Cold Comfort–Middle bays area, proposed that Speer's (1978) map was fundamentally correct in the Cold Comfort Bay portion and that Berg's (1975) map was the more accurate around Middle Bay. Schuh (1980b) completed a detailed stratigraphic examination of the unnamed formation, especially the iron formation and calcsilicate members.

Berg (1981) formally named the formation unconformably overlying the Upper Quartzite formation as the Falls Brook Group and defined the lowest member as a banded iron formation. He also suggested that correlations between the Mugford and Snyder groups are unjustified.

SULPHIDE MINERALIZATION

Speer (1972) described the Sulphide Hornfels (sic) as a massive sulphide-bearing unit interbedded with graphite hornfels and quartzite layers, in which the sulphide hornfels is thickest just south of Cold Comfort Bay, but the complete expression of the unit is obscured by large dykes.

In the type area on the shore of Cold Comfort Bay, Speer (1978) later suggested that the D formation, the Graphite–Sulphide Siltstone, is composed of finely laminated siltstone containing abundant pyrrhotite, pyrite and sphalerite (with exsolved chalcopyrite–presumably chalcopyrite 'disease'; cf. Barton and Bethke, 1987) in both disseminated and vein form. Two- to three-centimetre-thick sandy layers are interlayered with the siltstone. Speer *(op. cit.)* also found lenses of the Graphite–Sulphide Siltstone in the underlying Quartzite–Marble formation, and disseminated pyrrhotite, sphalerite and chalcopyrite in the Upper Quartzite formation.

Speer (1972) described the areal extent of the Upper Quartzite formation as being restricted to a hill south of Cold Comfort Bay. He observed that the unit was extensively mineralized by pyrite, pyrrhotite and minor chalcopyrite, and suggested that the sulphides were associated with a later intrusive body. He also stated that sulphide mineralization affecting the upper units of the Snyder Group were 'spatially associated' with the 'marginal intrusive rocks' (Wendy granodiorite).

Berg's (1975) revised map of the Snyder Group indicated that the Upper Quartzite formation and Graphite–Sulphide Siltstone formation extended much farther south to the Avakutakh River. Speer (1978) suggested that Berg actually mapped the Upper Quartzite formation as the Graphite–Sulphide Siltstone formation in the region between the second and third ponds south of Cold Comfort Bay, implying that the true Graphite–Sulphide Siltstone formation was restricted to the Cold Comfort Bay area, and that the Upper Quartzite formation contains sufficient sulphide mineralization to resemble the Graphite–Sulphide Siltstone formation. Berg (1975) also mapped two more outcrops of sulphide in Falls Brook, the presence of which were substantiated by Docka (1980).

Table 1. Geochemical data for Snyder Bay samples

Sample	SB91-004	SB91-006	SB91-007A	SB91-007B	SB91-007C	SB91-007D	SB91-009	SB91-010	SB91-016	SB91-017	SB91-018	SP91-020A
SiO_2	51.99	81.90	22.48	30.48	4.34	41.96	7.53	12.63	45.90	44.91	37.11	39.51
TiO_2	0.51	0.18	0.03	0.01	0.07	1.53	0.25	0.55	0.58	2.22	1.80	1.62
Al_20_3	7.21	9.10	0.32	0.12	0.61	11.01	1.30	3.18	9.89	12.80	13.64	10.67
Fe_20_3	1.54	0.27	5.63	4.31	5.44	1.83	5.31	4.63	1.89	1.47	1.61	1.42
FeO	12.47	2.20	45.56	34.90	44.02	14.82	42.97	37.51	15.29	11.92	13.05	11.46
MnO	0.04	0.01	0.26	0.14	0.01	0.09	0.03	0.06	0.05	0.14	0.23	0.19
MgO	2.36	1.00	3.71	1.09	0.00	6.15	0.00	1.01	1.75	10.20	5.05	5.49
CaO	2.37	2.25	5.34	3.84	0.39	5.10	1.42	3.75	5.77	5.87	11.28	9.13
Na_20	1.33	2.19	0.00	0.00	0.06	1.96	0.12	0.29	2.85	3.08	1.15	2.89
K_20	0.61	0.45	0.00	0.01	0.09	0.31	0.13	0.22	0.54	0.17	0.24	0.30
P_20_5	0.10	0.07	0.61	0.87	0.04	0.19	0.09	0.07	0.22	0.28	0.22	0.16
Cr	126.01	60	33	31	218	138	74	71	124	296	76	94
Ni	698.51	18	8	28	956	0	667	496	230	148	62	55
Sc	14.52	0	9	0	0	46	0	16	8	35	41	38
V	1675.95	104	63	20	791	432	481	646	388	461	466	642
Cu	418.87	18	33	183	1198	59	401	317	198	180	33	107
Pb	16.67	237	11	0	32	82	25	18	18	7	0	27
Zn	3588.87	84	233	31	2571	820	225	980	95	144	98	2173
S	74968.33	9730	18290	137841	265516	9008	314201	241176	87017	7791	1167	37408
As	0.00	0.00	0.00	0.00	33.00	0.00	346.19	186.71	16.46	0.00	0.00	44.87
K	5097.00	3760	17	100	722	2598	1046	1785	4508	1453	1976	2490
Rb	8.36	4	0	1	2	4	4	4	4	0	2	2
Ba	289.15	114.73	0.00	0.00	66.99	622.54	82.89	113.56	498.86	0.00	153.01	217.26
Sr	52.78	50.37	17.34	11.81	18.02	209.97	59.95	114.32	185.37	95.96	431.09	237.61
Ga	0.00	0.00	0.00	0.00	0.00	18.79	0.00	0.00	0.00	21.73	18.12	16.43
Nb	11.19	2.80	2.77	1.66	3.90	14.52	5.34	6.85	8.36	8.21	12.78	11.48
Zr	84.16	156.56	13.31	6.16	20.35	112.27	43.10	39.85	137.53	143.39	85.05	115.27
Ti	3033.00	1055	162	54	432	9190	1517	3285	3501	13315	10785	9718
Y	33.86	9.9	22.0	21.5	18.2	21.0	23.4	20.8	23.0	29.8	23.2	29.6
Th	4.97	0.0	0.0	0.0	0.0	0.0	4.7	0.0	0.0	0.0	0.0	0.0
U	31.31	4.1	0.0	0.0	22.6	0.0	17.7	7.9	3.4	0.0	4.0	4.2
Ce	39.91	0.0	0.0	61.1	0.0	82.2	44.3	50.7	42.2	65.0	43.6	99.6
Cl	35.10	49.02	70.18	94.89	0.00	41.39	24.90	20.37	38.24	27.38	76.15	49.29

* NR = not recorded; trace element values in ppm; major oxides in wt%

THE PRESENT STUDY

REGIONAL GEOLOGY

As the preceding compilation indicates, there is substantial variation in intrepretation as to where the sulphide-bearing formation occurs within the Snyder Group, including the areal extent of the formation and its relationships to intrusive rocks. In this study, the stratigraphy has not been examined in detail; instead, work has concentrated essentially on sampling and studying the sulphide-bearing formation within the Snyder Group as mapped by Berg (1975). The Falls Brook Group has not been examined because sulphide zones have not been reported within the unit. The present authors agree with Smyth and Knight's (1978) contention that the so-called iron formation within the Snyder Group is more properly termed an iron silicate formation; outcrops of true iron formation consisting of appreciable iron oxide were not found.

The authors agree with Berg's (1975) extension of the Graphite−Sulphide Siltstone and Upper Quartzite formations south through to the Avakutakh River. The consistency in stratigraphic position and relatively uniform thickness (up to 1 to 3 m) of the sulphide layer suggest that it clearly represents a mappable unit, rather than simply interbeds within the Upper Quartzite formation (as suggested by Speer, 1978). As a result of most of the previous studies being concentrated on stratigraphic examinations of the Snyder−Falls Brook groups, the definition and delineation of the minor intrusive rocks that cut the sequences have been inconsistent and conflicting. The most critical example is the granodioritic intrusion that occurs within the Upper Quartzite and Graphite−Sulphide Siltstone formations. In more recent reports, this intrusive unit has been referred to as the Border breccia (Speer, 1978) and the igneous breccia (Docka, 1980). Such appellations only add to the confusion because on the western side of Cold Comfort Bay, in the contact region between the Lower Quartzite and the Archean rocks, there is a very distinctive and totally unrelated unit called the Snyder breccia (Barton and Barton, 1975; Speer, 1978).

The Snyder breccia is a medium- to dark-grey intrusive rock that contains rounded to angular xenoliths of Snyder Group quartzite and Archean basement rocks. According to Speer (1978), the breccia is composed of up to 15 percent rounded to subrounded gneiss, angular amphibolite and biotite schist, and uncommon, angular Snyder Group quartzite xenoliths. Barton and Barton (1975) derived an Rb−Sr isotopic age of 1842 + 17 Ma for the matrix of the breccia and K−Ar ages of 1257 ± 9 and 1281 ± 9 Ma for biotite and hornblende

Table 1. (continued)

SB91-020B	SB91-020C	SB91-020D	SB91-025	SB91-026	SB91-030A	SB91-030B	SB91-030C	SB91-030D	MAYNARD-CM	MAYNARD-CR
39.14	10.07	5.64	57.27	61.83	48.05	50.23	52.21	50.77	0.00	0.00
0.19	0.58	0.28	0.41	0.50	0.67	0.88	0.77	0.86	0.00	0.00
1.89	2.07	1.11	7.33	19.51	13.65	19.08	19.65	19.40	0.00	0.00
3.66	4.64	5.41	1.62	0.04	0.86	0.62	0.71	0.75	0.00	0.00
29.66	37.57	43.79	13.10	0.30	6.94	5.01	5.76	6.06	0.00	0.00
0.03	0.04	0.05	0.03	0.00	0.04	0.03	0.05	0.03	0.04	0.04
0.03	0.25	0.00	1.86	1.55	2.25	2.35	3.35	2.67	0.00	0.00
1.79	0.78	1.06	1.57	0.31	2.26	1.17	2.40	1.17	0.00	0.00
0.60	0.27	0.13	1.14	4.13	1.37	3.85	3.17	3.83	0.94	0.36
0.18	1.29	0.16	2.74	6.40	1.92	3.50	1.07	3.47	3.13	3.37
0.03	0.03	0.03	0.10	0.03	0.02	0.02	0.01	0.02	0.00	0.00
101	190	160	160	123	289	188	221	169	170	61
0	642	941	985	0	8	105	386	186	154	40
10	18	11	13	16	23	14	16	16	0	0
1095	1378	986	1916	390	1810	307	594	360	610	280
114	477	1352	268	6	82	63	62	134	144	28
28	75	47	54	25	13	5	5	7	37	24
240	1990	3044	1495	0	110	121	101	106	101	129
22118	251766	328656	76496	1693	5142	15682	16724	24607	0	0
246.42	136.88	166.07	0.00	0.00	23.93	0.00	0.00	0.00	0.00	0.00
1494	10692	1287	22779	53145	15922	29046	8849	28789	0	0
2	20	4	61	137	41	30	11	31	0	0
79.91	451.90	94.87	632.21	1450.96	1381.21	930.79	326.00	1356.90	1290.00	6550.00
40.40	56.18	46.48	145.27	222.59	134.29	78.92	236.30	104.26	113.00	380.00
0.00	0.00	0.00	0.00	26.55	21.77	40.20	27.74	32.12	73.00	NR
2.96	3.77	3.63	11.23	13.05	5.44	7.79	8.51	8.95	0.00	0.00
45.17	45.28	26.42	78.30	144.04	126.28	156.64	141.41	133.31	230.00	110.00
1151	3483	1655	2458	2992	4005	5294	4592	5168	0	0
2.7	17.9	14.4	51.0	14.0	9.4	3.9	5.8	2.9	41.0	NR
0.0	0.0	0.0	4.6	12.8	4.6	0.0	0.0	0.0	0.0	0.0
6.0	32.0	17.7	46.3	5.2	0.0	0.0	0.0	0.0	0.0	0.0
0.0	41.3	36.2	64.4	66.5	37.2	65.6	110.2	88.6	0.0	0.0
51.60	59.27	22.87	92.38	28.27	41.44	27.81	28.15	10.67	0.00	0.00

separates from the breccia. They suggested that the 1842 Ma age was probably that of the breccia and the 1257-1281 Ma ages represent cooling ages for the Kiglapait Intrusion. Speer (1978) defined this breccia as a premetamorphic intrusion into the Snyder Group, with the metamorphic peak being contact metamorphism caused by the Kiglapait Intrusion. The authors could find no reaction or hornfels rims around any of the xenoliths, which, when coupled with the subrounded to round nature of some of the basement xenoliths, suggests that the breccia was an explosive, fluidized intrusion.

Speer (1978) describes the igneous breccia between the Snyder Group and the Outer Border Zone (Falls Brook Group of Berg, 1975) as a synmetamorphic intrusion related to the Kiglapait Intrusion as an anatectic melt. On the eastern shore of Cold Comfort Bay and in outcrop on the eastern shore of the third pond from the bay, this granodioritic intrusive contains angular xenoliths of Snyder Group sedimentary rocks and mafic granulite. In places near the third pond, the xenoliths have thin (3 to 5 mm thick) reaction rims composed of almandine. Unlike the xenoliths within the Snyder breccia, which exhibit fluidized intrusion-related shape modification and have no hornfelsic rims, the xenoliths in the granodiorite look like fragments stoped from a cooler country rock. This granodioritic rock is readily distinguishable from the Snyder breccia and thus it is suggested here that the name Wendy granodiorite be retained for this intrusive unit.

PRELIMINARY RESULTS

As indicated on Figure 1, sulphide zones were sampled at seven localities. Samples collected (Table 1) were grab samples, and where zones were sufficiently thick, a series of samples were collected across the outcrop width.

Site 1 comprises the two sulphide layers (samples 16 and 17) in Falls Brook. The thicker layer is 1.5 m wide and consists of massive, layered pyrrhotite and pyrite in silicious siltstone containing quartzite interlayers. The second layer to the south consists mainly of disseminated pyrrhotite and pyrite in a granular siltstone.

Site 2 is a sulphide zone (sample 18) between Falls Brook and Site 3. It is composed of massive pyrrhotite–pyrite sulphide lenses within graphitic siltstone–quartzite. The host rock also contains very fine-grained disseminated pyrite.

Site 3 is a sulphide zone (sample 20) on strike from that at Site 2. The mineralization occurs as lensoid, massive pyrrhotite containing pyrite hosted by siltstone. The sulphide zone bifurcates from a massive 6-m-thick layer to thinner lenses, 0.15 to 0.3 m thick, which enclose host-rock fragments. This bifurcation and engulfing of country rock is suggestive of ductile remobilization of sulphide during deformation. The outcrop of Site 3 is very heavily weathered to gossan, and the massive sulphides are friable.

Site 4 is the type locality (samples 25 and 26) for the Graphite–Sulphide Siltstone formation (after Speer, 1978). On the shore, the outcrop is a 1- to 2-m-wide interlayered siltstone, quartzite and sulphide sequence. The outcrop to the south of the beach consists of a 10-m-high gossan zone containing massive pyrrhotite and siltstone interlayers.

There is a ridge between Sites 3 and 4 extending from the point between Middle and Cold Comfort bays to the first pond. The ridge is underlain by Upper Quartzite formation outcrops that were extensively intruded by the Wendy granodiorite. Outcrop exposures of the quartzite and granodiorite are commonly rusty weathering due to the high contents of disseminated pyrite within both rock types.

Site 5 (sample 30) by the third pond, is a 3- to 4-m-thick siltstone and quartzite unit containing massive pyrrhotite–pyrite. It is complicated by the intrusion of numerous dykes of the Wendy granodiorite; the granodiorite itself is rusty weathering due to disseminated pyrite. The rusty granodiorite is very conspicuous and can be traced both downstream in the river from the third pond and on the ridge to the east of the pond.

Site 6 is a small, 4-m-across outcrop (sample 4) of very rusty, gritty siltstone. Pyrrhotite, including minor chalcopyrite, is present as thin (<1 cm) layers within the siltstone.

Site 7 (samples 6, 7, 9 and 10) was sampled over a lateral distance of more than 500 m. The rock types are interlayered siltstone and quartzite containing either massive pyrrhotite lenses up to 1 to 2 m thick, or disseminated pyrrhotite in host rocks. Pyrrhotite is also present as sheared, recrystallized pods containing xenoliths of host rock. Pyrite, and much less abundant chalcopyrite, occur in the host rocks as disseminations and microstringers. The massive pyrrhotite areas are very heavily stained and gossanous.

GEOCHEMICAL DATA

Major- and trace-element data for the sulphide zones are listed in Table 1. The analyses were completed on pressed-powder whole-rock pellets using a fully automated XRF at the Department of Earth Sciences, Memorial University of Newfoundland. Data reduction and plotting were executed using Clarke's (1991) NEWPET program. As indicated by data in Table 1, the base-metal contents of these sulphide zones are generally low and unremarkable. Copper contents range from 6 to 1352 ppm, Pb from 0 to 237 ppm, and Zn from 0 to 3589 ppm. Barium ranges from 0 to 1451 ppm and As contents are generally very low but may reach up to 346 ppm. The most enriched lense sampled was at Site 3, followed by Site 7–two areas where sulphide zones were not mapped by Speer (1978).

When compared to the average metal contents of D-Band ore, the lowest grade ore at the Sullivan deposit (Hamilton *et al.*, 1983), as listed in Table 2, the sulphide samples from the Snyder Group are of very low grades.

Table 2. Average major oxide (wt%) and base-metal concentrations (ppm) in D-Band ore, Sullivan Mine (from Hamilton *et al.*, 1983)

Fe_2O_3	33.61 %
MnO	0.315 %
Cu	330*
As	830*
Pb	13900
Zn	66800

* note Cu and As are average values for Sullivan ore

Based on Cu–Pb–Zn ratios within the samples (Figure 2), the Snyder Group sulphide zones would be classified as volcanogenic massive sulphide type rather than sedimentary exhalative (SEDEX) type (after Lydon, 1983) because of very low Pb contents. Only the sample having the highest Pb content (SB91-6 @ 237 ppm) would be classified as SEDEX type.

Maynard (1991) suggested that shale-hosted Pb–Zn–Ba deposits can be subdivided into two main types: (1) continent margin(CM)-type and (2) cratonic rift(CR)-type. The CM-type deposits are developed on the slope to basin floor of continental margins proximal to closing oceanic basins, and hence exhibit the geochemical influence of oceanic crust. They are generally base metal poor and Ba rich. The CR-type systems are developed within rifts in continental crust and are Pb–Zn rich. The CR-type deposits, therefore, are more economically important in that they may contain ore-grade concentrations of Pb and Zn.

Maynard (1991) compiled geochemical data from the literature and defined typical end-member compositions for host rocks to the CM- and CR-types of mineralization. The biggest differences between the two types lie in the Na–K ratios (higher in CM), Co–Ni–Cu contents (higher in CM) and Sr and Ba contents (both higher in CR). Maynard *(op. cit.)* deduced that these geochemical differences can be ascribed to different proportions of continental- and oceanic-crust involvement in the generation of the two types.

Figure 3 shows extended trace-element variation diagrams for the sulphide zones in the Snyder Group. In Figure 3a, the samples are ratioed to Maynard's (1991) CM end-member values and on Figure 3b to Maynard's CR end-member values. Also illustrated on each diagram are values for the other type (i.e., on Figure 3a, the values from CR are ratioed to CM, and vice versa on Figure 3b). These diagrams reveal striking element-distribution patterns for the Snyder Group samples in that the field for all samples

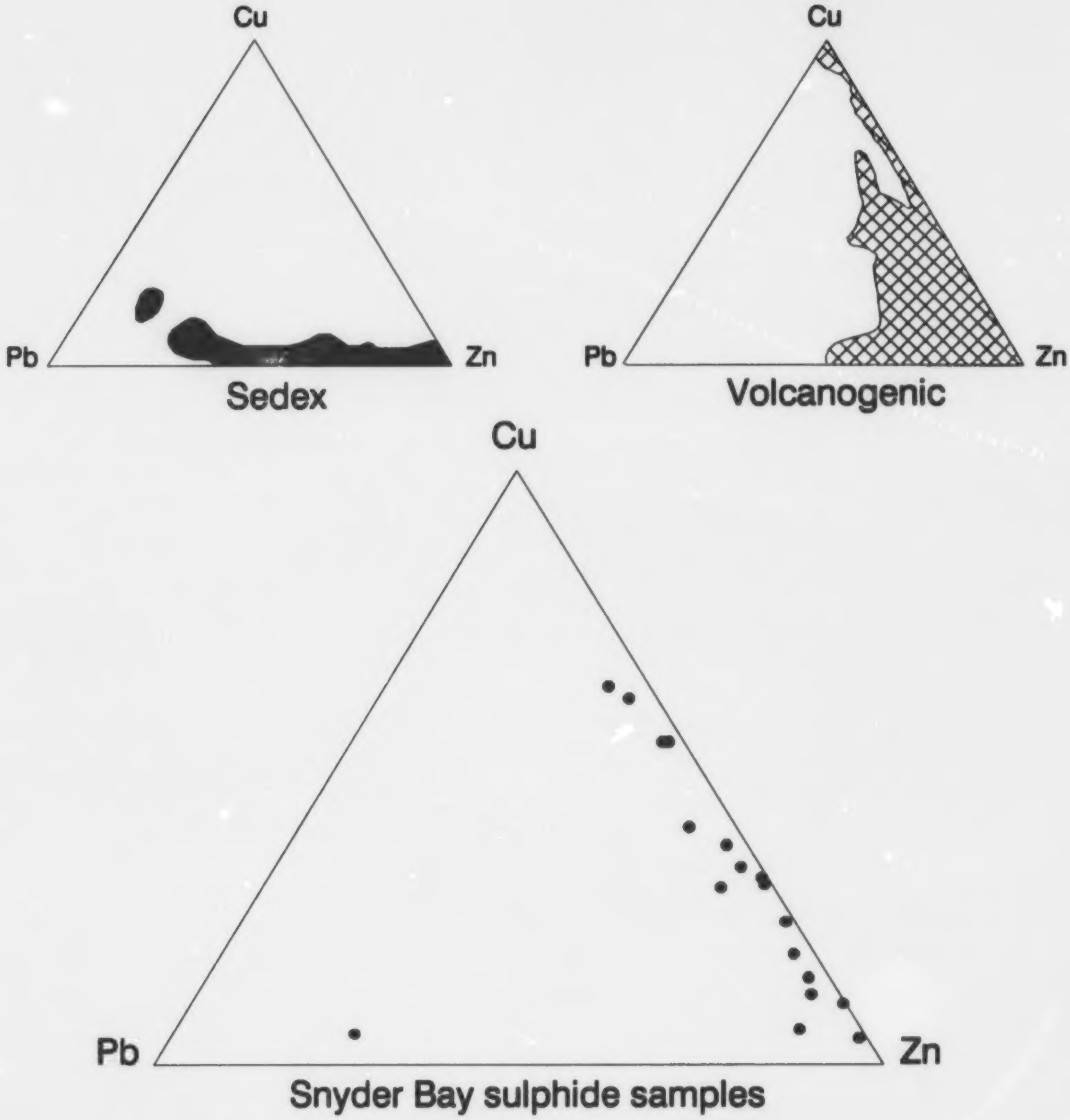

Figure 2. *Base-metal (Pb–Cu–Zn) variation diagram for samples from the sulphide zones in the Snyder Group. The inset diagrams show Lydon's (1983) fields for volcanogenic massive sulphide (VMS) deposits and sedimentary exhalative massive sulphide (SEDEX) deposits.*

parallels the element variations of the CM end member but is out of synchronicity with the variations in the CR end member. In particular, the average CM-type end member exhibits a similar Na_2O depletion relative to K_2O and MnO as the Snyder Group samples,and the concentrations and distributions of other elements, including Ni, Cu, Ba, Zn, V and Cr, are similar in the Snyder Group to those in the CM end member.

This close correlation of geochemical parameters suggests that the Graphite–Sulphide Siltstone formation of the Snyder Group was deposited in the Early Proterozoic on a continental margin with oceanic crust outboard. Such a supposition is at odds with Speer's (1978) suggestion that the Snyder Group was deposited as an epicontinental shelf sequence within the Archean craton. Definition of the sulphide zones in the Snyder Group as being similar to continental-margin mineralization, coupled with the low contents of base metals as determined in this study, unfortunately suggests that the possibility of locating economic concentrations of base metals within the Snyder Group is highly unlikely.

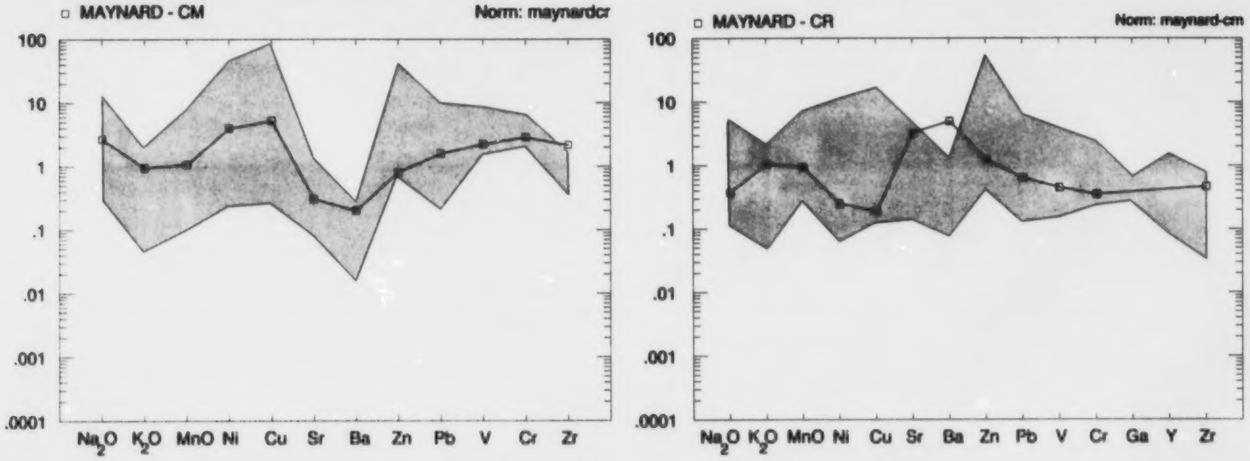

Figure 3. *(a) Normalized extended trace-element variation diagram for samples from the Snyder Group. The range in values for the Snyder Group samples is illustrated by the shaded field on the diagram, and the values for Maynard's (1991) average host rocks to Cratonic Rift-type Pb–Zn–Ba mineralization are denoted by the open symbols. The normalization factors are Maynard's (1991) average host-rock values for Continental Margin-type Pb–Zn–Ba mineralization. (b) Normalized extended trace-element variation diagram for samples from the Snyder Group. The range in values for the Snyder Group samples is illustrated by the shaded field on the diagram, and the values for Maynard (1991) average host rock to Continental Margin-type Pb–Zn–Ba mineralization are denoted by the open symbols. The normalization factors are Maynard's (1991) average host-rock values for Cratonic Rift-type Pb–Zn–Ba mineralization.*

CONCLUSIONS

The Snyder Group consists of five sedimentary formations that have been contact metamorphosed by the Kiglapait Intrusion. One of these formations, the Graphite–Sulphide Siltstone formation, contains massive sulphide lenses and sulphide disseminations. The sulphides are predominantly pyrrhotite and lesser pyrite, and minor chalcopyrite. Sphalerite has been reported by previous workers and although not positively identified in hand specimens from this study, the geochemical data indicate elevated Zn contents in some samples. The base-metal contents of the sulphide zones are rather low, certainly well below grade necessary for economic exploitation. The relative apportionments and absolute contents of trace and major elements suggest that the sulphide zones in the Snyder Group are analogous to those that form on continental margins, and hence they exhibit the geochemical influence of oceanic crust.

Granodioritic intrusions within the unit contain abundant disseminated pyrite and consequently have rusty-weathering outcrops. These rusty granodiorites can be readily mistaken for the similarly rusty-weathering and spatially close sedimentary sulphide zones.

ACKNOWLEDGMENTS

The efforts of R. Goudie of Enterprise Newfoundland and Labrador, R. Wardle and B. Greene of the Newfoundland Department of Mines and Energy (NDME), and B. Cox of the Office of Research, Memorial University (MUN) in setting up this research contract are very gratefully acknowledged. B. Ryan of the NDME and R. Emslie of the Geological Survey of Canada (GSC) are thanked for providing interesting discussions on the geology of northern Labrador, for sharing their accommodations and good cheer in Nain, and their helicopter during a very wet and miserable mid July on the coast. D. Clarke completed the XRF analyses. Ms. E. Hearn constructed the diagrams from a great distance, supervised the analytical and preparation work, and acted as forensic accountant for this research grant. R. Wardle is thanked for continued fruitful discussion on the Proterozoic supracrustals of Labrador.

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NEWFOUNDLAND COMPUTERIZES ITS CLAIMS-RECORDING OFFICE

N. Gover
Mineral Rights, Mineral Lands Division
Mineral Resource Management Branch

INTRODUCTION

Management of Newfoundland's Ground-Staked Exploration Licences has entered the 'Computer Age'. In March of 1991, the final phase of the development and implementation of a computerized Mineral Rights Ground-Staked Licencing System was completed. This computerized claims-recording and licence-management system was developed by Newfoundland and Labrador Computer Services Limited from specifications prepared by Newfoundland's Mineral Rights Recorder.

SYSTEM GENERAL

The Mineral Rights System is a networked application allowing simultaneous use by several users. As installed, the Mineral Rights System runs on a Novell ELS Level II Network. It is a microcomputer system using a Dell 310 PC as a file server, and four other Dell PC's as work stations. The file server has a 90 Mb hard drive with 4 Mb of RAM and is used in the non-dedicated mode. All of the work stations have hard drives ranging from 20 to 40 Mb, consequently, they may be used as stand alone units.

SYSTEM STRUCTURE

The Mineral Rights System is relatively large, with respect to the software image size and the number of database and index files accessed, hence the system is partitioned into three smaller subsystems as follows:

A. Main Subsystem
B. Licencing Subsystem
C. Report Production Subsystem

Also, the Backup, Restore and Archive utilities are separated out and are performed using SYTOS batch files. The SYTOS software itself is used in conjunction with the TLC LEGACY tape backup unit to perform these functions. The SYTOS software is relatively large (over 140K) and thus it could not be incorporated into one of the three Mineral Rights subsystems.

To link all the subsystems and utilities together, the Novell menu system is used with a Mineral Rights System template. As a bonus, a Novell Utilities sub-menu has also been included whereby the user can perform such functions as monitoring the LAN print queues, obtain a LAN logon list, send messages to other users logged on to the LAN, and execute DOS commands. For more information on the Novell Utilities the user can consult the relevant Novell documentation.

Figure 1 represents a flow chart of the system.

SOFTWARE

The program is written in dBase III (Ashton-Tate) and compiled with the Clipper Compiler (Nantucket). The program is compiled and this allows it to be used independent of the dBase software.

PURPOSE

The system provides fast and efficient storage, retrieval and management of the ground-staked licence records. Exploration statistics and various reports are produced in order to chart the trends in the mineral exploration industry to enable better resource management.

In addition, manual preparation of time-consuming correspondence has been eliminated. Preparation of licences and covering letters, acknowledgment of receipt of applications, cancellation notices for the licence holder, and display notices have been automated.

SYSTEM OPERATING HIGHLIGHTS

In Newfoundland, a ground-staked mineral licence is comprised of any number of claims and claim blocks. A claim is a 400 m^2 and a claim block is any number of claims staked in a square or rectangular group, up to and including 64 claims. A prerequisite to staking is the purchase and use of claim tags. The system maintains a list of claim tags that have been sold and the tag status, which could be active, used or returned.

Upon receipt of an application for a licence, all relevant information to create a licence record is entered into the system, such as name, location and staking particulars. The system will assign a correspondence file number, a licence number, the recording date, the issuance date and the work report due date (Screen L17).

A file of statutory holidays is maintained to ensure that the assigned dates are working days only. In addition, a letter of acknowledgment for the applicant, a mailing label and central registry correspondence file-folder label are generated and printed (Screen L61).

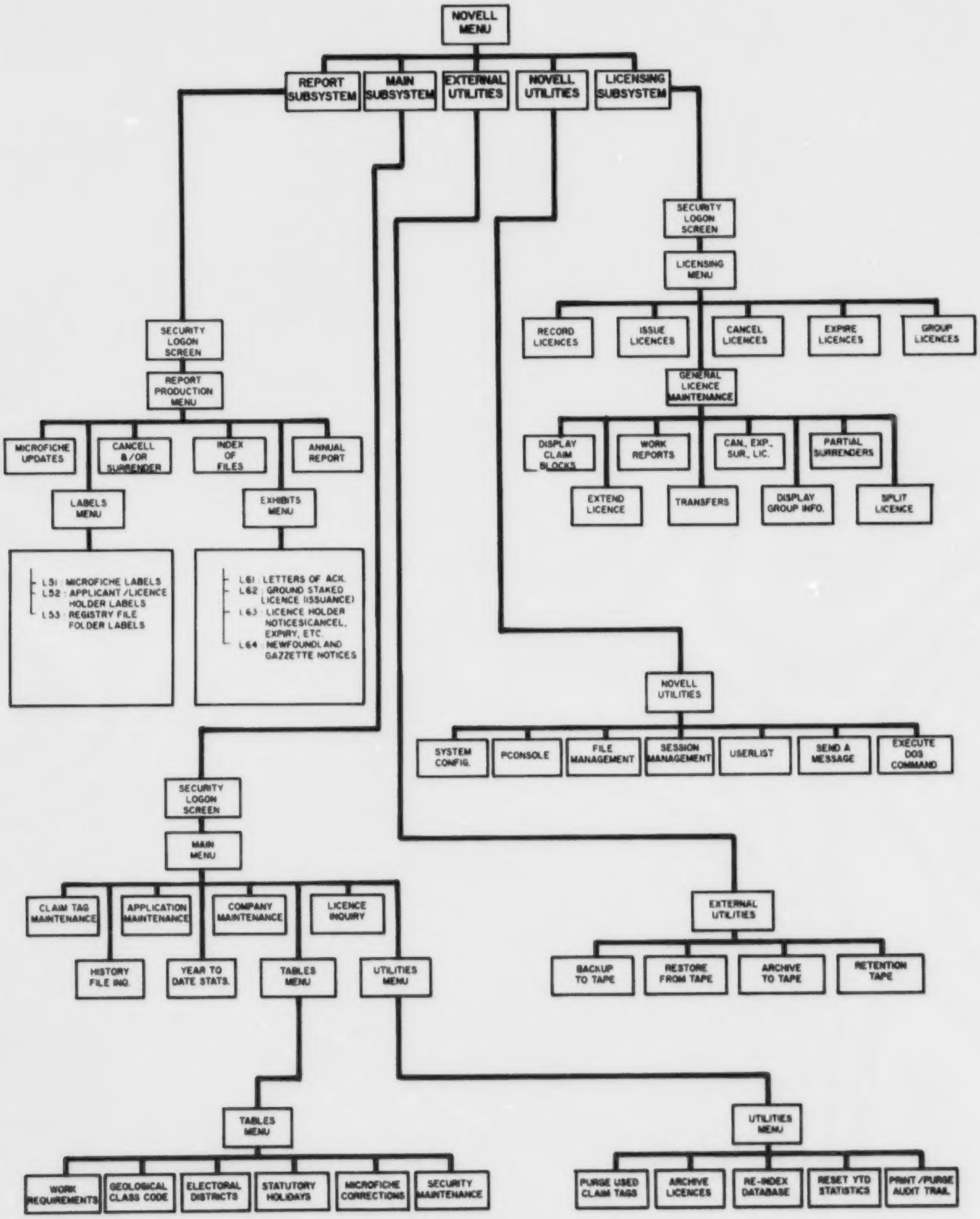
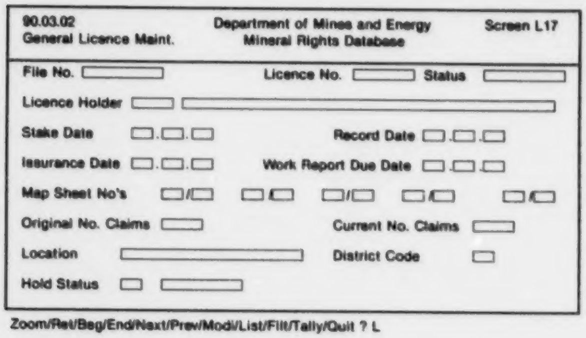

NOVELL MENU
REPORT SUBSYSTEM
MAIN SUBSYSTEM
EXTERNAL UTILITIES
NOVELL UTILITIES
LICENSING SUBSYSTEM
SECURITY LOGON SCREEN
LICENSING MENU
RECORD LICENCES
ISSUE LICENCES
CANCEL LICENCES
EXPIRE LICENCES
GROUP LICENCES
GENERAL LICENCE MAINTENANCE
DISPLAY CLAIM BLOCKS
WORK REPORTS
CAN., EXP., SUR., LIC.
PARTIAL SURRENDERS
EXTEND LICENCE
TRANSFERS
DISPLAY GROUP INFO.
SPLIT LICENCE
SECURITY LOGON SCREEN
REPORT PRODUCTION MENU
MICROFICHE UPDATES
CANCELL &/OR SURRENDER
INDEX OF FILES
ANNUAL REPORT
LABELS MENU
EXHIBITS MENU
L51 : MICROFICHE LABELS
L52 : APPLICANT /LICENCE HOLDER LABELS
L53 : REGISTRY FILE FOLDER LABELS
L61 : LETTERS OF ACK.
L62 : GROUND STAKED LICENCE (ISSUANCE)
L63 : LICENCE HOLDER NOTICES(CANCEL, EXPIRY, ETC.
L64 : NEWFOUNDLAND GAZZETTE NOTICES
NOVELL UTILITIES
SYSTEM CONFIG.
PCONSOLE
FILE MANAGEMENT
SESSION MANAGEMENT
USERLIST
SEND A MESSAGE
EXECUTE DOS COMMAND
SECURITY LOGON SCREEN
MAIN MENU
CLAIM TAG MAINTENANCE
APPLICATION MAINTENANCE
COMPANY MAINTENANCE
LICENCE INQUIRY
HISTORY FILE INQ.
YEAR TO DATE STATS.
TABLES MENU
UTILITIES MENU
EXTERNAL UTILITIES
BACKUP TO TAPE
RESTORE FROM TAPE
ARCHIVE TO TAPE
RETENTION TAPE
TABLES MENU
WORK REQUIREMENTS
GEOLOGICAL CLASS CODE
ELECTORAL DISTRICTS
STATUTORY HOLIDAYS
MICROFICHE CORRECTIONS
SECURITY MAINTENANCE
UTILITIES MENU
PURGE USED CLAIM TAGS
ARCHIVE LICENCES
RE-INDEX DATABASE
RESET YTD STATISTICS
PRINT /PURGE AUDIT TRAIL

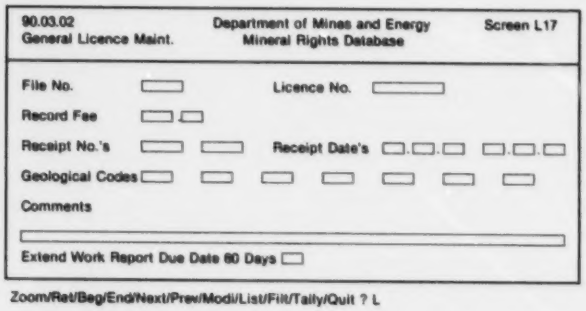
Figure 1. *Schematic of the Mineral Rights System.*



90.03.02 Department of Mines and Energy Screen L17
General Licence Maint. Mineral Rights Database

File No. Licence No. Status
Licence Holder
Stake Date Record Date
Issurance Date Work Report Due Date
Map Sheet No's
Original No. Claims Current No. Claims
Location District Code
Hold Status

Zoom/Ret/Beg/End/Next/Prev/Modi/List/Filt/Tally/Quit ? L

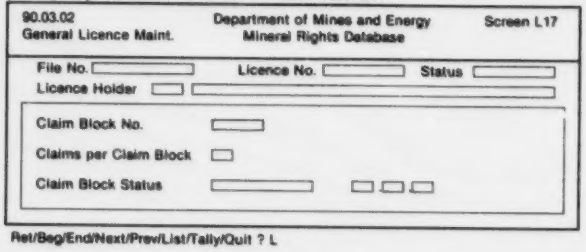


90.03.02 Department of Mines and Energy Screen L17
General Licence Maint. Mineral Rights Database

File No. Licence No.
Record Fee
Receipt No.'s Receipt Date's
Geological Codes
Comments
Extend Work Report Due Date 60 Days

Zoom/Ret/Beg/End/Next/Prev/Modi/List/Filt/Tally/Quit ? L

Display Claim Block Information



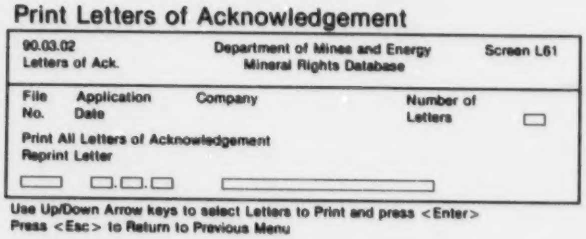
90.03.02 Department of Mines and Energy Screen L17
General Licence Maint. Mineral Rights Database

File No. Licence No. Status
Licence Holder
Claim Block No.
Claims per Claim Block
Claim Block Status

Ret/Beg/End/Next/Prev/List/Tally/Quit ? L

Screen L17

Print Letters of Acknowledgement



90.03.02 Department of Mines and Energy Screen L61
Letters of Ack. Mineral Rights Database

File No. Application Date Company Number of Letters
Print All Letters of Acknowledgement
Reprint Letter

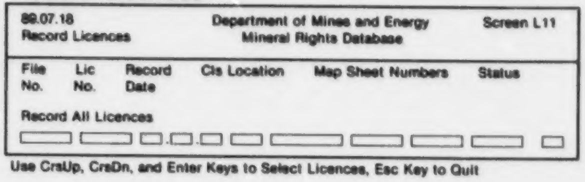
Use Up/Down Arrow keys to select Letters to Print and press <Enter>
Press <Esc> to Return to Previous Menu

Screen L61

In order to avoid typographical errors in the licence holder name and for convenience of entering, all names are maintained in a company file. Therefore, whenever it is necessary to enter a name, such as for new applications, issuance of new claim tags or transfer of licences, the name is available from the company file by accessing it via a pop-up menu and automatically assigning it to the record.

The regulations require that applications shall not be accepted (recorded) until after the expiry of 30 days from the date of staking. Also, licences cannot be issued until after the expiry of 30 days from the date of acceptance. Therefore, the system is prompted daily for recording and issuing dates (Screen L11 and 12).

Record Licences

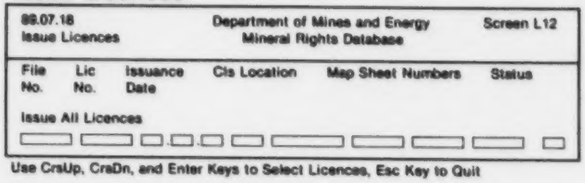


89.07.18 Department of Mines and Energy Screen L11
Record Licences Mineral Rights Database

File No. Lic No. Record Date Cls Location Map Sheet Numbers Status
Record All Licences

Use CrsUp, CrsDn, and Enter Keys to Select Licences, Esc Key to Quit

Issue Licences



89.07.18 Department of Mines and Energy Screen L12
Issue Licences Mineral Rights Database

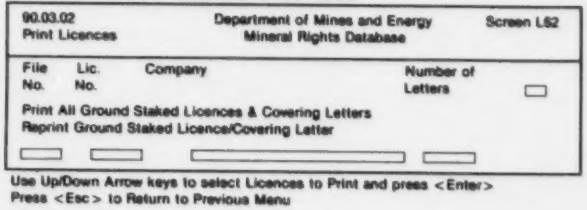
File No. Lic No. Issuance Date Cls Location Map Sheet Numbers Status
Issue All Licences

Use CrsUp, CrsDn, and Enter Keys to Select Licences, Esc Key to Quit

Screen L11 and 12

Licences are recorded and issued by a simple stroke of the 'ENTER' key. The licence itself is computer generated, and printed together with a covering letter for the licence holder (Screen L62).

Print Ground Staked Licences



90.03.02 Department of Mines and Energy Screen L62
Print Licences Mineral Rights Database

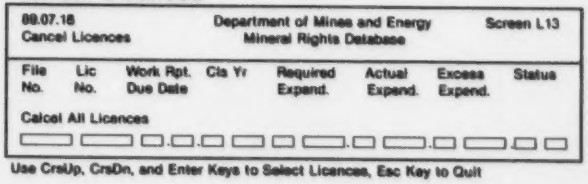
File No. Lic. No. Company Number of Letters
Print All Ground Staked Licences & Covering Letters
Reprint Ground Staked Licence/Covering Letter

Use Up/Down Arrow keys to select Licences to Print and press <Enter>
Press <Esc> to Return to Previous Menu

Screen L62

In a similar way, work report due dates are checked and licences that are deficient in assessment work credit may be cancelled automatically (Screen L13).

Cancel Licences



89.07.18 Department of Mines and Energy Screen L13
Cancel Licences Mineral Rights Database

File No. Lic No. Work Rpt. Due Date Cls Yr Required Expend. Actual Expend. Excess Expend. Status
Calcel All Licences

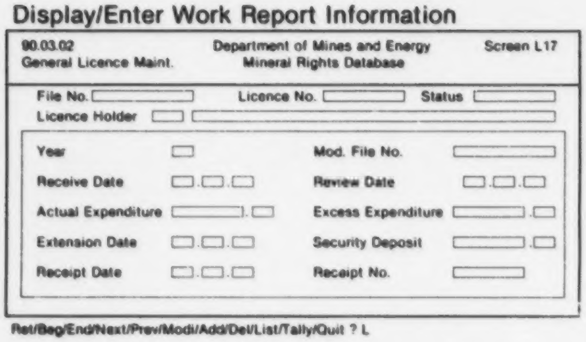
Use CrsUp, CrsDn, and Enter Keys to Select Licences, Esc Key to Quit

Screen L13

Licences with sufficient excess assessment work credit for the next year are highlighted so that the excess credit may be carried forward. Also, the work report due date is automatically moved up a year.

When licences are cancelled, a notice for the licence holder informing him of the cancellation and the reason for the cancellation is generated and printed. An appropriate display notice declaring the lands covered by the cancelled licence to be open for staking at a fixed date is also produced. Similarly, expiry dates are checked and licences that have expired are noted and the appropriate correspondence is printed.

Yearly assessment work calculations are automatic. For example, as assessment work reports are accepted, it is a simple matter to retrieve the licence or licences covered by the report and enter the date the report was received, reviewed and the value of the assessment work expended. The required value of the work for the year covered by the report, and any excess work credit is computed by the system and the record is updated. Yearly work requirements per claim are stored by the system to facilitate this computation (Screen L17).


Display/Enter Work Report Information

90.03.02
General Licence Maint.
Department of Mines and Energy
Mineral Rights Database
Screen L17
File No.
Licence No.
Status
Licence Holder
Year
Mod. File No.
Receive Date
Review Date
Actual Expenditure
Excess Expenditure
Extension Date
Security Deposit
Receipt Date
Receipt No.

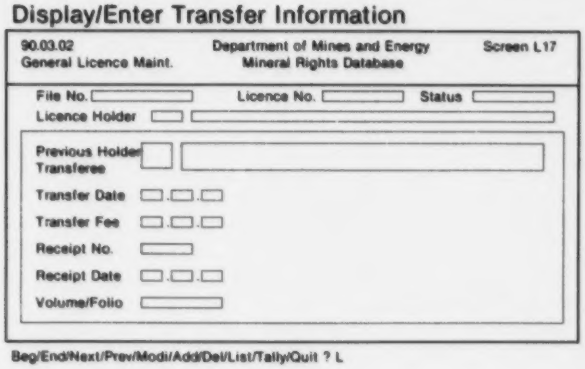
Ret/Beg/End/Next/Prev/Modi/Add/Del/List/Tally/Quit ? L

Screen L17

Newfoundland regulations allow for the grouping of two or more coterminous licences, provided all of the licences to be grouped have passed their first anniversary date and are in good standing. To group two or more licences, it is only necessary to enter the date of the grouping and the licence numbers to be grouped. The computer produces a number for the new grouped licence. A complete new grouped licence record is generated, which includes the number of claims and any excess work credits calculated from the grouped licences. In addition, the new licence and a covering letter for the licence holder is automatically printed.

A licence holder may surrender a part of any licence at any time by submitting a notice of surrender. Complete claim blocks or portions of claim blocks may be surrendered by restaking the area(s) to be kept. Areas to be retained may be coterminous or consist of one or more separate areas. In the latter case, new reduced licences are issued. All combinations of partial surrenders are easily handled by the system. The Department need only highlight the claim blocks being surrendered, add the new claim blocks and assign any new licences with the appropriate blocks. In a similar manner to grouping, complete new records are generated by the system, with new reduced licences and covering letters being printed. Also, display notices for the blocks being surrendered are printed.

The system will allow records to be noted when licences have been transferred. We simply retrieve the licence, zoom to the transfer screen and add the transferee, transfer date, transfer fee, receipt number and date, and the volume and folio numbers that indicate the filing location of the transfer document (Screen L17).


Display/Enter Transfer Information

90.03.02
General Licence Maint.
Department of Mines and Energy
Mineral Rights Database
Screen L17
File No.
Licence No.
Status
Licence Holder
Previous Holder
Transferee
Transfer Date
Transfer Fee
Receipt No.
Receipt Date
Volume/Folio

Beg/End/Next/Prev/Modi/Add/Del/List/Tally/Quit ? L

Screen L17

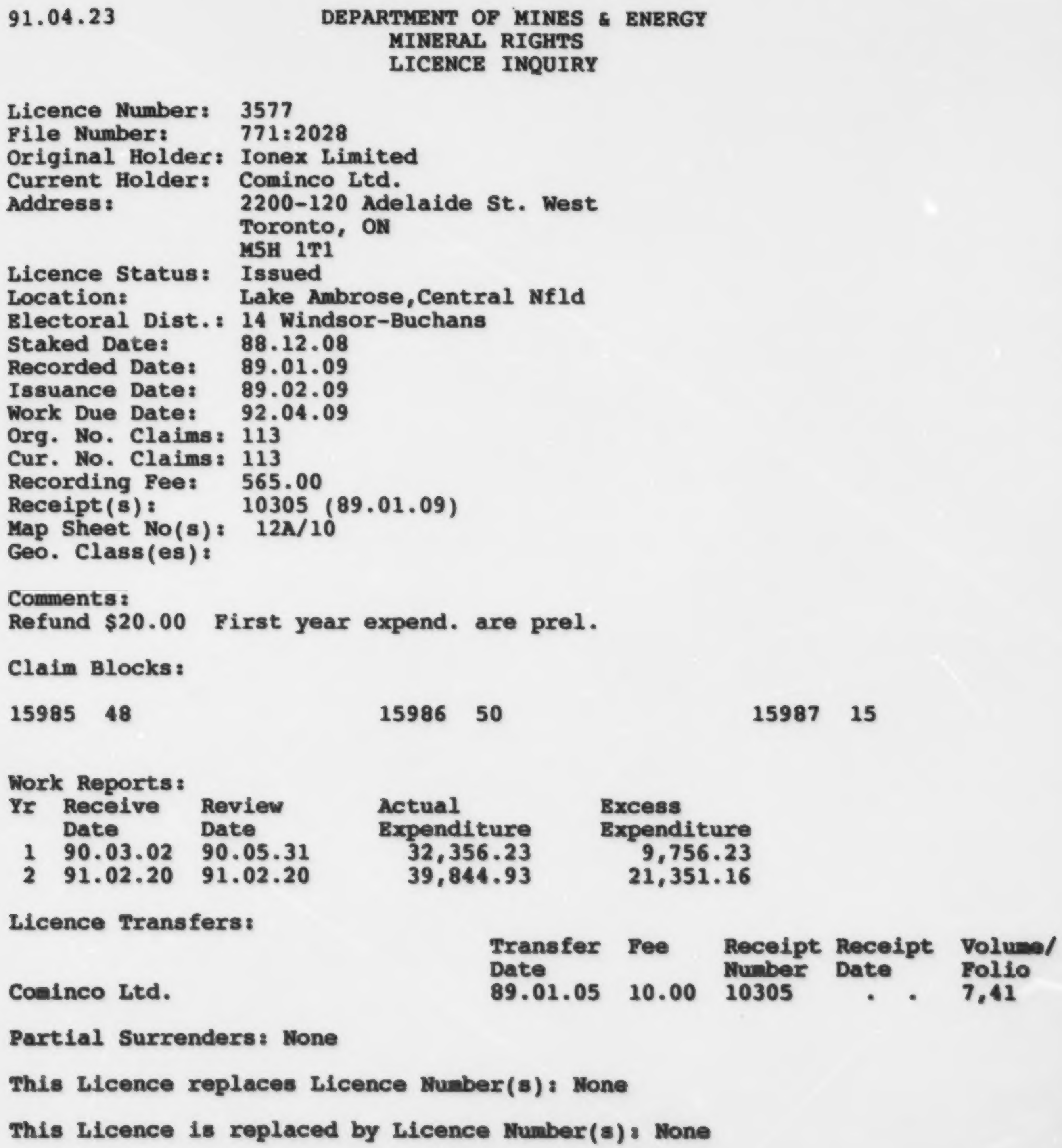
The retrieval of licence records is as easy as entering a licence number or a claim block number. A complete record relating to the licence (block) is available for viewing or printing. See Figure 2 for a sample printout.

Report production is an integral part of the system. Several standard reports are generated and printed when needed. The three main ones being a monthly microfiche update report, which consists of a record of all changes and additions to the database for every month. A quarterly cancellation/surrenders/expiry report, which contains a listing of all licences surrendered, cancelled, partially surrendered and expired for any given quarter, and an annual report, containing a listing of all licences issued for a given year (Screen L21 and L24).

It is possible to sort through the main database and filter out selected data. For example, we can list all licences held by a particular company, all licences issued during a given period, all licences contained within a particular map sheet, all licences at a given location, and combinations. For example, all licences issued to company 'X' during period 'Y' for a given map sheet.

Current year to date statistics respecting number of claims staked, number of claims cancelled/surrendered/expired, and total number of claims in good standing are available. The statistics are also available for any year back to 1980.

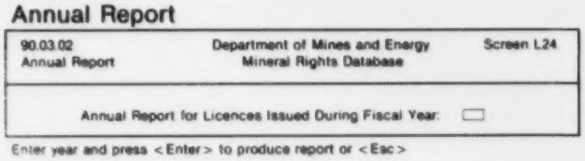
A Backup and Restore utility is provided in order to facilitate system security. The system is backed up daily using a three tape rotation, and a monthly backup using one tape. Security of the system is further enhanced by providing each user with a logon id and a password for both the Network and the Mineral Rights Database.



```
91.04.23                 DEPARTMENT OF MINES & ENERGY
                               MINERAL RIGHTS
                              LICENCE INQUIRY

Licence Number:    3577
File Number:       771:2028
Original Holder:   Ionex Limited
Current Holder:    Cominco Ltd.
Address:           2200-120 Adelaide St. West
                   Toronto, ON
                   M5H 1T1
Licence Status:    Issued
Location:          Lake Ambrose,Central Nfld
Electoral Dist.:   14 Windsor-Buchans
Staked Date:       88.12.08
Recorded Date:     89.01.09
Issuance Date:     89.02.09
Work Due Date:     92.04.09
Org. No. Claims:   113
Cur. No. Claims:   113
Recording Fee:     565.00
Receipt(s):        10305 (89.01.09)
Map Sheet No(s):    12A/10
Geo. Class(es):

Comments:
Refund $20.00  First year expend. are prel.

Claim Blocks:

15985  48                       15986  50                       15987  15


Work Reports:
Yr  Receive   Review        Actual             Excess
    Date      Date          Expenditure        Expenditure
 1  90.03.02  90.05.31         32,356.23           9,756.23
 2  91.02.20  91.02.20         39,844.93          21,351.16

Licence Transfers:
                                          Transfer  Fee    Receipt  Receipt  Volume/
                                          Date             Number   Date     Folio
Cominco Ltd.                              89.01.05  10.00  10305     .  .    7,41

Partial Surrenders: None

This Licence replaces Licence Number(s): None

This Licence is replaced by Licence Number(s): None
```

Figure 2. *An example of a sample printout.*

Annual Report

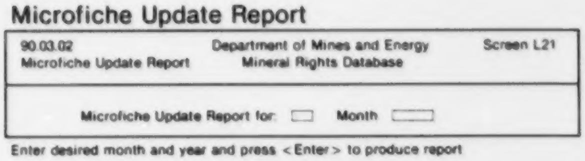


90.03.02 Department of Mines and Energy Screen L24
Annual Report Mineral Rights Database

Annual Report for Licences Issued During Fiscal Year: ☐

Enter year and press <Enter> to produce report or <Esc>

Microfiche Update Report



90.03.02 Department of Mines and Energy Screen L21
Microfiche Update Report Mineral Rights Database

Microfiche Update Report for: ☐ Month ☐

Enter desired month and year and press <Enter> to produce report

Screen L21 and 24

SUMMARY

The highly competitive and fast moving nature of the mineral exploration industry dictates that any recording office be able to respond to the industries needs in an efficient and timely manner. Our computerized system allows us to process claim information and compile statistics faster, thereby allowing us to respond promptly to the queries of the industry. Also, this tool allows the Department to perform its task of administering Newfoundland's mineral regulations fairly and effectively.

CONTACTS

Further information is available from:

Noel F. Gover, Mineral Claims Recorder
Mineral Lands Division
Department of Mines and Energy
P.O. Box 8700
St. John's, Newfoundland
A1B 4J6

Telephone: 709-729-2773
Fax: 709-729-6782

*Current Research (1992) Newfoundland Department of Mines and Energy,
Geological Survey Branch, Report 92-1, pages 447-451*

CORE-STORAGE PROGRAM, 1991

A. Harris and S. Cochrane
Mineral Lands Division
Mineral Resource Management Branch

INTRODUCTION

In 1978, the Newfoundland Department of Mines and Energy began a program of drill-core collection and storage. Initially, three core-storage libraries, having a combined capacity of 260,000 m of core samples, were established at Pasadena, St. John's, and Goose Bay. Funding for construction of these libraries was provided under the Canada–Newfoundland Mineral Development Subsidiary Agreement (1977-82) and construction was completed in 1982.

Since 1982, three additional core-storage libraries have been established at Springdale, Buchans and Baie Verte (Figure 1) increasing the combined core-storage capacity to a total of 785,000 m for all six core libraries. The core-storage libraries are operated by the Mineral Resource Management Branch of the Department of Mines and Energy and as of November 1991, housed a combined collection of approximately 715,891 m of cores (Figure 1).

DRILL-CORE ACQUISITION

The first several years of the core-collection program were devoted to identifying and collecting core samples from abandoned drilling sites. Collection of core samples from abandoned sites was completed in insular Newfoundland in 1981 and in Labrador in 1983, with an overall recovery rate of approximately 50 percent of core drilled. Core samples from this source form most of the present collection. All core samples from these sites have been catalogued and are available for inspection by the mineral-exploration industry or other interested persons.

New core acquisition by the core-storage program generally falls into one of the following categories:

(a) Core samples collected from recently abandoned mineral-exploration sites.

(b) Core samples donated to the program by mining and mineral-exploration companies.

Generally, complete drillhole sections are collected and where cores are accessible by road, all available samples are retrieved. In remote bush locations, complete drillhole sections are collected but only from representative drillholes (Plate 2).

1991 CORE COLLECTION

During 1991, drill-core acquisition continued in insular Newfoundland. The core-storage program acquired 24,164 m of core samples during this past year (Table 1).

Drill-core collection is continuing this fall and it is anticipated that a further 51,000 m of core samples will be delivered to our core-storage facilities by mineral-exploration companies this year. These core samples will be catalogued during the 1992 field season and made available for inspection at that time.

A catalogue of all open-core samples in Department core libraries is available from the Publications and Information Section, Geological Survey Branch of the Department of Mines and Energy.

USING THE CORE LIBRARIES

Any person who wishes to visit any of the core libraries for the purpose of examining core samples should provide advance notice to the person in charge of the particular core library. The Pasadena core library is staffed on a full-time basis by one employee and is open to the public during regular government office hours. Advance notice of one day is required in order to properly accommodate visitors to the Pasadena core library. Visitors to the St. John's core library should also provide advance notice of one day and advance notice of one week is required to properly accommodate visitors to the Goose Bay core library (Plate 3) because it is not staffed on a full-time basis. Visits to the Springdale, Buchans and Baie Verte core libraries require a minimum notice of two days.

The indexing system in all three core libraries is based on the National Topographic System (NTS). Core samples from each drillhole are assigned a unique master number based on the 1:50 000-scale topographic sheet on which the drillhole collar is located. A manual index card-file forms the basis of this system in conjunction with a 1:50 000-scale drillhole location map file, which shows the location of all drillholes. A completed example of the index card is shown in Appendix I. Each core library also contains a file of all available data, i.e., drillhole logs, cross-sections, and assay results for all core samples stored. The company drillhole number and the assessment file in which the drilling was reported are referenced on the index card. The manual index card-file and the map index are supplemented by a

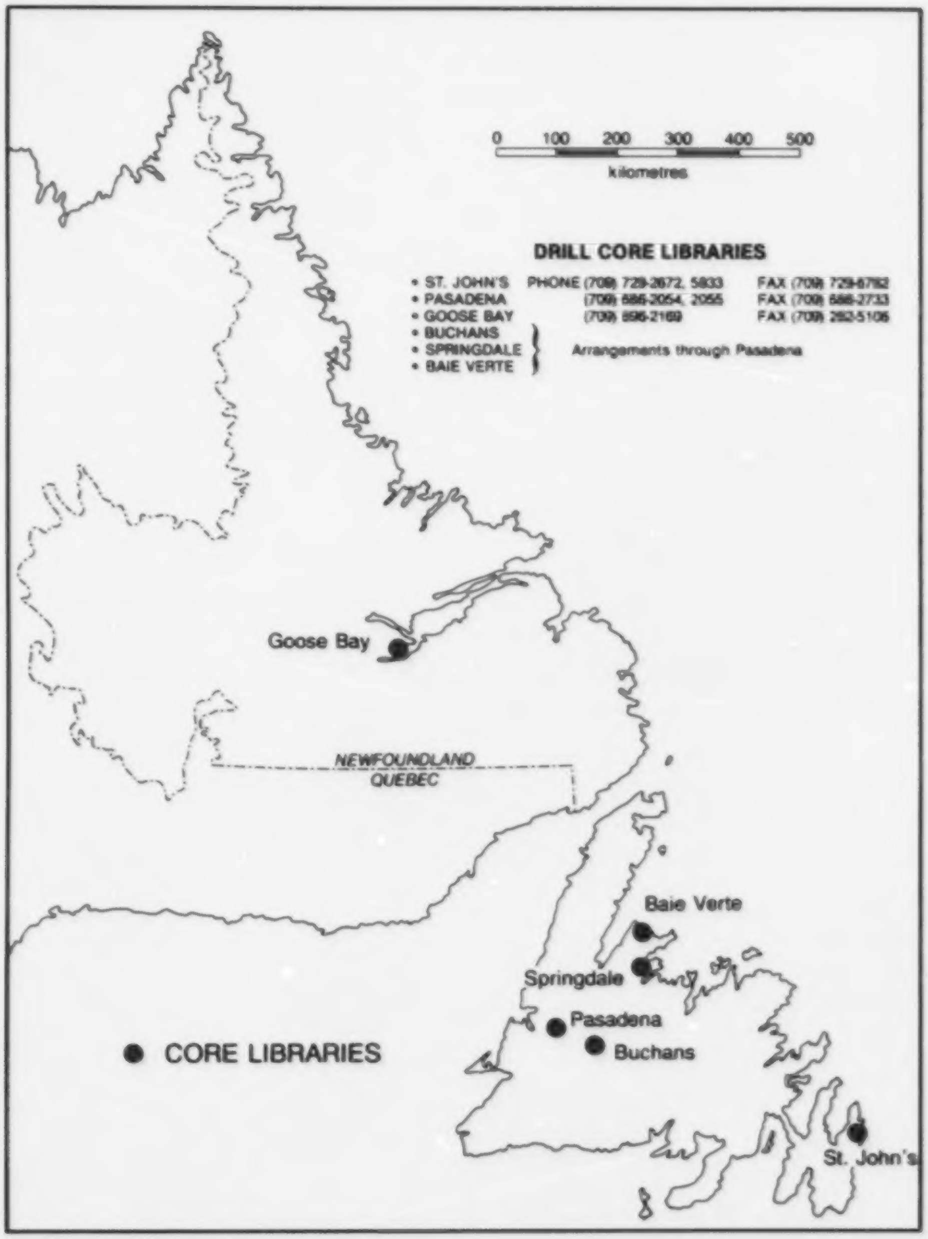

0 100 200 300 400 500
kilometres
DRILL CORE LIBRARIES
• ST. JOHN'S PHONE (709) 729-2672, 5833 FAX (709) 729-6782
• PASADENA (709) 686-2054, 2055 FAX (709) 686-2733
• GOOSE BAY (709) 896-2169 FAX (709) 282-5106
• BUCHANS
• SPRINGDALE
• BAIE VERTE
Arrangements through Pasadena
Goose Bay
NEWFOUNDLAND
QUEBEC
Baie Verte
Springdale
Pasadena
Buchans
St. John's
● CORE LIBRARIES

Figure 1. *Core libraries in Newfoundland and Labrador.*

computerized drillhole index, which contains all of the information on the previous two indexes. The computer drillhole index-file functions as a permanent master-file, from which catalogues of all of our drill-core holdings are updated yearly and printed for distribution at the Department's Annual Review of Activities each November. The file is resident on microcomputers in the Department's St. John's office and in the Pasadena core library, where it is available for production of customized indexes.

All six core libraries have core examination areas that contain or share, with other core libraries, rock-cutting equipment, a core splitter, a stereomicroscope, a magnetic susceptibility meter, a resistivity meter, an ultraviolet light and a McPhar TV-1 scintillometer.

Sampling is generally permitted where doing so does not destroy any lithological sequence or leave gaps in the drillhole record. The smallest size sample that is useful is taken and the minimum amount of core that must be left in each core box is a one-quarter-size split of the original core. The user must complete a *'Request To Sample Form'* (Appendix II) before sampling is permitted. All core samples, pulps, powders, thin sections, and any other materials generated

Table 1. Core collected in 1991.

Property	Company/Year	No. Holes	Metres
Betts Cove	(Inco) 1988-90	35	4,860
Baie Verte	(Corona) 1988-90	97	11,224
Bobby's Pond	(Inco) 1991	2	1,237
Avalon Peninsula	(Inco) 1990	2	346
Hickey's Pond	(Corona) 1990	4	420
Burlington Road	(Terra Gold) 1989	2	275
Jackson's Arm	(Carrick Resources) 1990	17	1,336
Cape Ray	(Dolphin Exploration) 1989	29	3,616
Goose Arm	(Len Pye) 1990-91	4	122
Wellsdale Fee Simple	(Wellsdale Mnrls) 1979	18	728

Plate 1. *Pasadena core library.*

Plate 2. *Core collection from remote bush location.*

from the core samples must be returned (along with a copy of the results of any work done on the samples) to the core library within one year from the date the samples were taken. The user shall be responsible for costs incurred in returning samples to the core library.

There are no fees charged for the services provided by the core-storage program, however, patrons making extensive use of the trim saws to sample drill core are required to supply their own blades or replace the blades that are used.

Plate 3. *Goose Bay core library.*

The core-storage program does not have a practice for core reduction or discarding of 'redundant' core.

FURTHER INFORMATION

Potential users of the core-storage program should contact the people listed below to arrange visits to the core libraries and to obtain information on the core-storage program.

Project Geologist–Core-Storage Unit (Alvin Harris)
Department of Mines and Energy
P.O. Box 8700
St. John's, Newfoundland
A1C 5T7
PHONE: (709) 729-5833
FAX: (709) 729-6782

Core-Storage Geologist (Pasadena) (Stewart Cochrane)
Department of Mines Energy
Dr. A.K. Snelgrove Core Library
Pasadena, Newfoundland
A0L 1K0
PHONE: (709) 686-2054, 2055
FAX: (709) 686-2733

APPENDIX I

MASTER NO.: 1L/13 0007 UTM: 584100 5204900 21

COMPANY: Nfld. Dept. of Mines & Energy CO. NO.: F-7 DRILLING DATE: 66-12-19

MIN NO.: PROPERTY: Fortune STATUS:

DATE STORED: 82-02-25 LOG AVAIL.: Yes ASSAYS AVAIL.: Yes MDD NO.: 1M/4 (87)

CORE STORAGE INDEX CARD

INTERVAL (m)	BOXES	CORE	RACK LOCATION FROM	RACK LOCATION TO	MISSING INTERVALS (m)
0-48.76	6	AX	01B04A09	01B04B02	34.13-38.10
55.77-89.30	4	EX	01B04B03	01B04B06	48.76-55.77
E.O.H.					

CORE EXAMINATION FREQUENCY

81	82	83	84	85	86	87	88	89	
90	91	92	93	94	95	96	97	98	

APPENDIX II

CORE STORAGE UNIT

Request to Sample Core

USER NAME:

COMPANY/AFFILIATION:

ADDRESS:

NATURE OF PROJECT:

WORK TO BE CONDUCTED ON SAMPLES:

DRILL HOLE NOS:

SAMPLE INTERVALS:

SIZE AND SHAPE OF SAMPLES:

REQUEST DENIED/GRANTED (REASON):

DATE CORE STORAGE GEOLOGIST

NOTE: All core samples and/or pulps, powders and thin sections, etc., generated from core samples are to be returned to core library at the end of a previously specified period. A copy of all assays and other analytical work conducted on the samples is also required at the end of this period.

RETURN DATE: SIGNATURE:

*Current Research (1992) Newfoundland Department of Mines and Energy,
Geological Survey Branch, Report 92-1, page 452*

GEOSCAN–THE NATIONAL DATABASE FOR GEOLOGICAL INFORMATION: THE NEWFOUNDLAND CONTRIBUTION

Catherine Patey and Janet Gillespie
Publications and Information Section

GEOSCAN is a computerized bibliographic database of Canadian-produced and/or Canada-related geoscience documents. The GEOSCAN database and related indexes are used daily by industry, the general public and by callers to the Section, to identify documents on the geoscience of this province and the rest of Canada.

The Department of Mines and Energy, Government of Newfoundland and Labrador, is one of fourteen agencies that index documents into this national database. The departmental files, including mineral-assessment reports, mineral-inventory cards, published reports and maps, university theses, etc., are the sources of the indexed records. Additional documents, produced and edited by related agencies, e.g., the Geological Survey of Canada, are the source of other indexed records referring to this province.

Since 1982, mainly mineral-assessment file records have been added to the database. According to the results of a survey, in February, 1991, of the database and our collection of geoscientific documents on this province (e.g., Geofiles), 90 percent of the assessment files are in the database. In fact, over 80 percent of the documents in the Geofiles are referenced in GEOSCAN. This summer, basic bibliographic records for the remaining university theses in the Geofiles were entered, so that now all the theses i.e., all 349 in the Geofiles, are represented.

As of November, 1991, there were 127,500 records in GEOSCAN, including 11,000 referring to this province. The 126,500 non-confidential records are available through on-line computer searches at the Geological Survey of Canada Library in Ottawa, at the Publications and Information Section of the Geological Survey Branch in St. John's and at the 12 other indexing agencies across Canada. The number of locally requested searches has increased each year with over 80 completed so far this year. In order to improve in-house access to the information, a user-friendly subset of GEOSCAN for the province was introduced and is available for searching by staff and visitors at the Publications and Information Section of the Geological Survey Branch.

Information on documents referring to the geoscience of this province are also available from various printed GEOSCAN Newfoundland and Labrador indexes:

1. NTS (National Topographic System), Keyword, and Author (both corporate and personal) indexes to documents (including those authored by the Geological Survey of Canada) referring to this province with records in the database up to February 1982.

2. Computer printout sorted by NTS, with references to mineral-assessment file records entered into the database during the period 1982-1984.

3. NTS Index (with cross-reference lists) to mineral-assessment file records received by the Publications and Information Section during the periods:

   –April 1985 to June 1988 (272 records)
   –July 1988 to June 1989 (Open File Nfld/1927) (628 records)
   –July 1989 to June 1990 (Open File Nfld/2068) (292 records)

A similar NTS index to mineral-assessment files, received during the period July 1990 to June 1991, will be available by the end of the year. As well as these larger printed indexes, smaller custom indexes, e.g., 'Peat in Newfoundland' are available for sale. Other computer printout indexes, e.g., 'Author Index to Published Documents', are available for in-house use.

Besides indexing and keying new records and modifying existing records, improved keying and editing procedures are being developed with WordPerfect macros, and new coding sheets are being produced. The authority files and thesauri used in coding are being enlarged, and all the terms used are being more clearly defined. These terms are maintained on various card-files and binders, e.g., company name card-file, mineral lands information card-file, drillhole information binder, etc. The information on these card-files and binders is being transferred to several mini databases on CDS/ISIS and Reflex II software for easier access and reference. These files and mini databases, which were designed to make indexing consistent, are also useful for in-house inquiries.

Suggestions for improvements are welcome and notification of any errors or discrepancies in the earlier indexes would be appreciated. For further information on GEOSCAN, please contact Catherine Patey (709) 729-6441.

Current Research (1992) Newfoundland Department of Mines and Energy,
Geological Survey Branch, Report 92-1, pages 453-457

MINERAL OCCURRENCE DATA SYSTEM

G. Stapleton, C.F. O'Driscoll, W.K. Parsons[1], J. Smith and D. King
Mineral Deposits Section

ABSTRACT

The Mineral Occurrence Data System (MODS) is a three-part database consisting of a manual Mineral Inventory File, mineral occurrence maps on geological bases and a computerized Mineral Inventory Database. The MODS contains information on approximately 4000 mineral occurrences and is designed to offer fast easy access to information on the province's mineral resources.

INTRODUCTION

The Mineral Occurrence Data System (O'Driscoll *et al.*, 1991) consists of a manual Mineral Inventory File, mineral occurrence maps on geological bases and a computerized database. The project is designed to offer an efficient information service on all the mineral occurrences in Newfoundland and Labrador (Figure 1).

Most of the MODS information was compiled from a systematic search of mineral-exploration company-assessment files. Other sources of information include publications of the Newfoundland Department of Mines and Energy, the Geological Survey of Canada, news items from the press, publications in geological and mining journals, and personal communications from mining company and government personnel.

MANUAL MINERAL INVENTORY FILE

The manual Mineral Inventory File is part of the National Mineral Inventory and consists of mineral occurrence reports that summarize all data on known mineral occurrences in the province. The file presently contains approximately 4000 reports, including all mineral occurrences for the Island of Newfoundland (Figure 2) and selected areas in Labrador (Figure 3). The file and topographic location maps have been microfiched and duplicates are available upon request.

The Mineral Inventory File is continually being updated. This file was started in 1978, and since then, much new geological mapping and exploration have been carried out and many new occurrences have been discovered. Updates were done in the past only on land tenure and major mineral occurrences. Areas that have been updated with new mineral occurrence descriptions include all of 12A, and major parts of 12B and 12H map areas. In addition, industrial minerals for all of Newfoundland have been documented.

Mineral occurrence maps having updated geological bases have been published at a 1:250 000 scale and are available upon request. In addition, selected areas have been published at 1:50 000 and 1:100 000 scales. These maps contain locations, a listing and a brief description of the occurrences. An industrial minerals map of the Island of Newfoundland at 1:1 000 000 scale on a geological base, is now available. This map contains locations, a listing and a brief description of occurrences.

COMPUTERIZED MINERAL INVENTORY DATABASE

Since 1978, a computerized Mineral Inventory Database, which parallels the manual file, has been developed and maintained. This database contains information selectively extracted from the manual file and presently resides on a Hewlett Packard 9000/560 Unix-operated minicomputer, located at the Geological Survey Branch, Department of Mines and Energy, 95 Bonaventure Avenue, St. John's. The Geological Retrieval and Synopsis Program (GRASP) developed by the United States Geological Survey (Bowen and Botbol, 1975) is used to manage and manipulate the database. Although GRASP is a powerful program that allows complex searches to be performed, it is not user friendly or available for microcomputers. In the past, in order for users to access the computerized Mineral Inventory Database a visit to the Geological Survey Branch offices and the help of MODS personnel was required.

DEVELOPMENT OF MODS/PC

With the proliferation of microcomputers and the need for easier access to MODS information, it was decided to develop a microcomputer version of this information system in 1990 (Stapleton *et al.*, 1991; Stapleton and Parsons, 1991). The Mineral Occurrence Data System microcomputer

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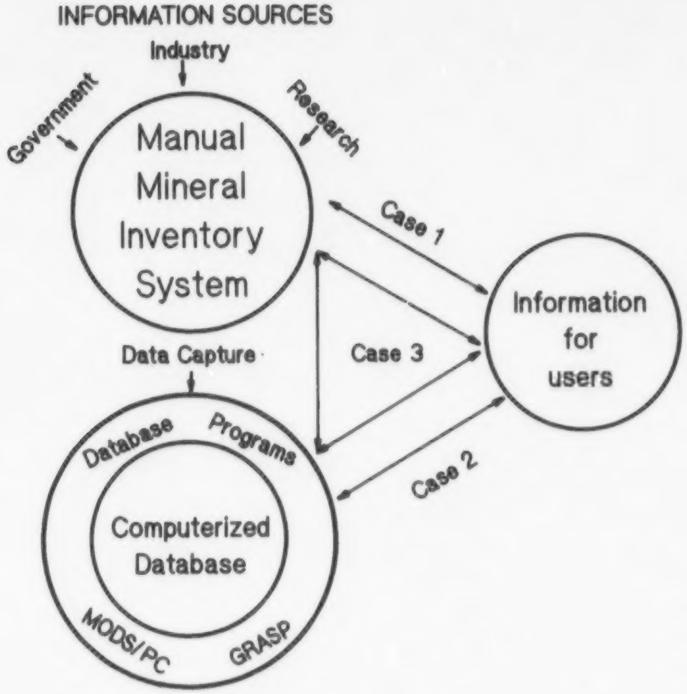
[1] Geoscience Data Centre

How MODS is Used


INFORMATION SOURCES
Industry
Government
Research
Manual Mineral Inventory System
Case 1
Information for users
Case 3
Data Capture
Database
Programs
Case 2
Computerized Database
MODS/PC
GRASP

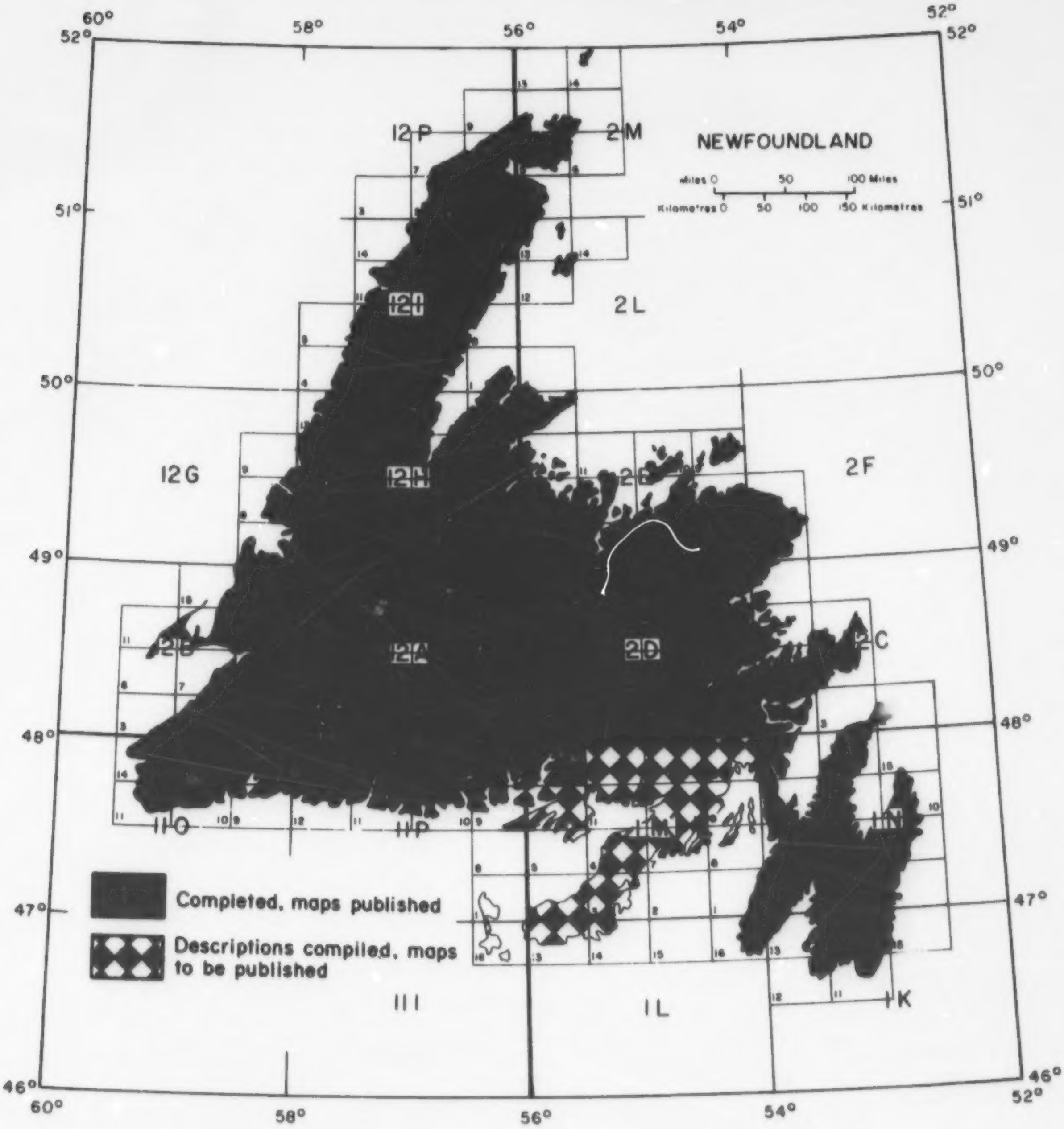
The Mineral Occurrence Data System can be used in three ways.

Case 1: The user wants detailed information on a specific mineral occurrence or group of mineral occurrences in a specific area. This information can be retrieved quickly using the manual mineral inventory file.

Case 2: The user wants less detailed information on all occurrences of a particular type, ie. vein type, located over a large geographic area. The computerized mineral deposits database can provide this information in printout format along with a computer generated location map.

Case 3: The user wants detailed comprehensive information on all occurrences of several types over a large geographic area. The purpose being to access the regions mineral potential and hopefully zero in on an area for field investigation and staking. In this case the computerized mineral deposits database is searched and used to narrow down and define specific occurrences or areas of interest. The more detailed manual mineral inventory reports can then be provided. The user can then use references contained in the manual reports to locate other information sources.

Figure 1. *Sketch illustrating how MODS is used.*


NEWFOUNDLAND
Miles 0 50 100 Miles
Kilometres 0 50 100 150 Kilometres
60°
58°
56°
54°
52°
52°
51°
50°
49°
48°
47°
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12P
2M
12I
2L
12G
12H
2E
2F
12B
12A
2D
2C
11O
11P
1M
1N
11I
1L
1K
Completed, maps published
Descriptions compiled, maps to be published

Figure 2. *Index map for Mineral Occurrence Data System project, insular Newfoundland.*

Application (MODS/PC V1.0) was written in-house by W. Keith Parsons, Systems Analyst with the Geological Survey Branch of the Department of Mines and Energy. It was written in the R:Base Database language (V2.11) and compiled using the R:Base compiler (V1.02). R:Base was chosen as the development tool because of the program's support of long string fields and the availability of a vendor-supplied compiler.

The minimum system requirements needed to run the MODS/PC program include an IBM-compatible microcomputer with an 80286 processor, 640 K of RAM, a hard disk, and a CGA videocard/monitor. As well, the system must have at least 475 K of free memory. To achieve this, certain memory resident programs may have to be deactivated.

The MODS/PC system is distributed on four 5 1/4" diskettes in a compressed format. An installation program (install.exe) aids the user in setting up the system by checking the target microcomputer's parameters, copying files to the

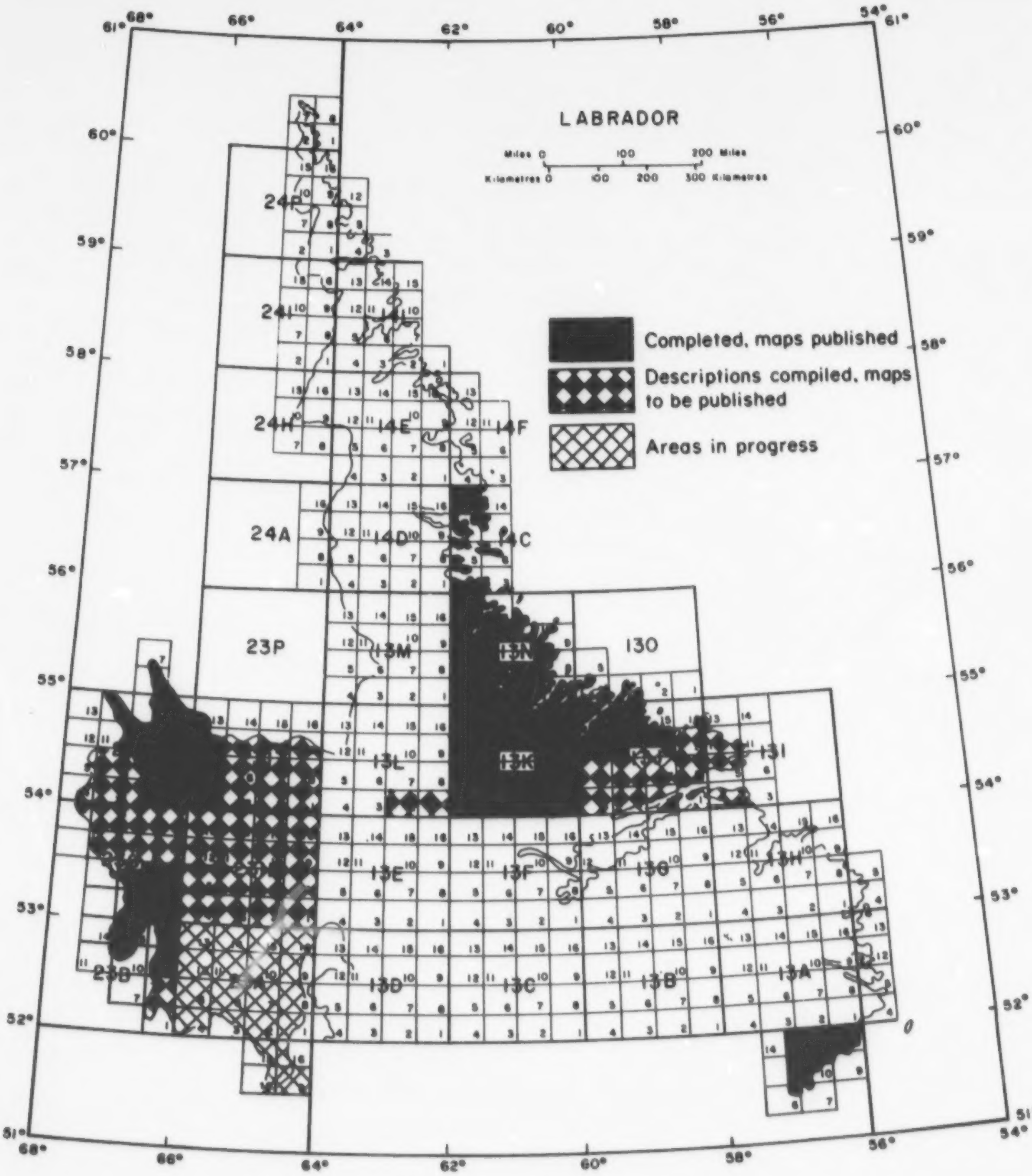
LABRADOR
Miles 0 100 200 Miles
Kilometres 0 100 200 300 Kilometres
Completed, maps published
Descriptions compiled, maps to be published
Areas in progress
68°
66°
64°
62°
60°
58°
56°
54°
61°
60°
59°
58°
57°
56°
55°
54°
53°
52°
51°
24P
24I
24H
24A
23P
23B
14E
14F
14D
14C
13M
13N
13O
13L
13K
13J
13I
13E
13F
13G
13H
13D
13C
13B
13A

Figure 3. *Index map for Mineral Occurrence Data System project, Labrador.*

hard disk, and decompressing these files. PKZIP, the industry standard compression utility distributed by PKware Inc., is used to compress the MODS/PC system, however, possession of this software is unnecessary as the files are in a self-extracting executable format. The compression ratio is better than ten to one so considerable hard-disk space is required.

Only one dataset may be installed at a time. The user may select between Newfoundland occurrences (2566 records occupying 6.1 M), Labrador occurrences (1259 records occupying 3.0 M) or a demonstration dataset (164 records occupying 420 kb). The program itself occupies about 400 kb of hard-disk space.

The MODS/PC program is menu-driven and provides the user with complete access to the computerized Mineral Occurrence Data System. The program allows the user to search, view and output occurrences from the database. Custom-designed searches are achieved by setting conditions and using Boolean logic to combine these conditions in different ways. Data portability is provided by the ability to write data to an ASCII delimited file. This allows input of a user-selected dataset to other application programs such as spreadsheets and databases or portability to other types of computer systems such as non-DOS microcomputers, minicomputers, or mainframes. Full user documentation is supplied with the program.

Each mineral deposit is described by a single MODS record, which has fifty-nine information fields. More detailed information on the deposit can be obtained by consulting key references listed in each record or by examining the hard-copy mineral inventory reports maintained at the Geological Survey Branch. These reports will allow information to be traced back to its source publication.

Version 1.0 of the MODS/PC system was released to the public in November, 1991. Periodically, upgrades to the system will be made available to the user. Subsequent releases will update the current records, add new occurrences, and provide revisions to the program. The MODS/PC program will be upgraded to version 3.1 of R:Base once a new compiler is released by Microrim Inc. The revisions will produce a modernized user interface, faster access times, and improved report generation. The R:Base compiler produces object files compatible with common 3rd-generation languages such as C. Future versions of MODS/PC may make use of this ability to add graphical options such as plotting and mapping to the program.

In future, when optical disk technology becomes widely used, it is planned to release all of the MODS reports on optical disk. This will enable subscribers to have the entire Mineral Occurrence Data System, including maps, MODS/PC and reports, in-house.

MODS USERS

The MODS is used primarily by mineral-exploration-company personnel, however, it is also used by mineral-exploration consultants, geotechnical consultants, personnel and students of academic organizations and the general public.

The MODS is used daily by government geologists in land-use planning. Advice is given to various departments of government in establishing wilderness areas, hydro developments, provincial and national parks, and any other developments that may conflict with future mineral exploration and development. In addition, municipal councils and the Department of Municipal Affairs are advised of the location, extent, and nature of mineral deposits in specific areas, so that new housing and commercial developments, municipal parks, water reservoirs and sewage-disposal systems can be located, where possible, in areas of low mineral potential.

Copies of the file are made available to the various agencies of the federal government such as the Mineral Policy Sector and the Geological Survey of Canada. These are then adapted to the National Mineral Inventory, MINSYS and CANMINDEX.

REFERENCES

Bowen, R.W. and Botbol, J.M.
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1991: A mineral occurrence database for microcomputers. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 91-1, pages 395-397.

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1991: MODS/PC—Mineral Occurrence Data System for personal computers, users manual. Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File Nfld. 2174, 80 pages and 4 disks.

Preliminary report on the metamorphic geology of the Port aux Basques Complex, southwestern Newfoundland[1]

Jerry L. Burgess[2], Michael Brown[2], and Cees R. van Staal
Continental Geoscience Division

Burgess, J.L., Brown, M., and van Staal, C.R., 1992: Preliminary report on the metamorphic geology of the Port aux Basques Complex, southwestern Newfoundland; in *Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 145-154.*

Abstract

This report summarizes initial results of a project concerned with unravelling the pressure-temperature-time-deformation history of the Port aux Basques Complex (PaBC). The Complex occurs between the Cape Ray Fault Zone and Rose Blanche, representing part of the Central Mobile Belt, and comprises the Grand Bay, Port aux Basques and Otter Bay divisions, separated by the Grand Bay Lineament and the Isle aux Morts Fault Zone, respectively. Rocks of all three divisions have been affected by polyphase deformation under variable metamorphic conditions. Structural analysis of the Port aux Basques Division reveals four phases of deformation, the first two of which occurred during high-grade metamorphism resulting in multiple stages of porphyroblast growth and matrix recrystallization. The metamorphism, which is of kyanite-sillimanite type and included development of migmatites by anatexis, is interpreted to be the result of crustal thickening as manifested in an overall clockwise pressure-temperature path.

Résumé

Le présent rapport résume les résultats préliminaires d'un projet visant à reconstituer l'évolution de la déformation selon le temps, la pression et la température du complexe de Port aux Basques (PaBC), qui se trouve entre la zone de failles de Cape Ray et Rose-Blanche; il fait donc partie de la zone mobile centrale et englobe les divisions de Grand Bay, de Port-aux-Basques et d'Otter Bay. Les deux premières divisions sont séparées par le linéament de Grand Bay tandis que les deux dernières le sont par la zone de failles d'Isle-aux-Morts. Les roches des trois divisions ont subi une déformation polyphasée dans des conditions métamorphiques variables. L'analyse structurale de la division de Port-aux-Basques met en lumière quatre phases de déformation, les deux premières ayant eu lieu pendant un épisode de métamorphisme intense; ce fait explique la croissance en plusieurs étapes des porphyroblastes et la recristallisation de la matrice. Le faciès métamorphique à kyanite et à sillimanite ainsi que la formation de migmatites par anataxie sont interprétés comme résultant d'un épaississement crustal, sur considération des variations dextrorsum d'ensemble de la pression et de la température.

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[1] Contribution to the Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-94

[2] Department of Geology, University of Maryland at College Park, MD 20742, USA

A new look at Gander–Dunnage relations in the Carmanville map area, Newfoundland

K.L. Currie
Continental Geoscience Division

Currie, K.L., 1992: A new look at Gander-Dunnage relations in Carmanville map area Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 27-33.

Abstract

The structural map of the Gander Zone (Cuff Pond pelite), lithologically and chemically resembles the base of the Dunnage Zone (Davidsville Group), but is more complexly deformed. Plutonic-metamorphic belts intersecting on the Gander River Ultramafic Belt suggest that thermal activity accompanied and locally outlasted deformation related to ductile shearing separating the Gander and Dunnage zones. Within the Dunnage Zone, the Hamilton Sound sequence can be matched unit by unit with the Exploits Group to the west. Early thrusts probably separate the Hamilton Sound sequence from the Davidsville Group. Disposition of sedimentary sequences and plutonic belts suggest presence of two or more major southerly-directed thrust sheets within the Dunnage Zone, with thermal activity concentrated along frontal and lateral ramps.

Résumé

Le sommet structural de la zone de Gander (pélite de Cuff Pond) ressemble lithologiquement et chimiquement à la base de la zone de Dunnage (groupe de Davidsville), mais est déformée de façon plus complexe. La présence de zones métamorphiques et plutoniques se recoupant dans la zone ultramafique de Gander River porte à croire qu'une activité thermique a accompagné la déformation liée au cisaillement ductile qui sépare les zones de Gander et de Dunnage, et qu'elle s'est poursuivie, par endroits, après cette déformation. A l'intérieur de la zone de Dunnage, la séquence de Hamilton Sound se laisse corréler unité par unité avec le groupe d'Exploits à l'ouest. Les failles de chevauchement plus anciennes séparent probablement la séquence de Hamilton Sound du groupe de Davidsville. La disposition des séquences sédimentaires et celle des zones plutoniques semblent indiquer la présence d'au moins deux grandes nappes de charriage de direction sud dans la zone de Dunnage, l'activité thermique étant concentrée sur le bord de rampes frontales et latérales.

Preliminary report on the structural control of the Rendell-Jackman gold deposit, Springdale Peninsula, Newfoundland[1,2]

B. Dubé, K. Lauzière, and D. Gaboury[3]
Quebec Geoscience Centre, Quebec

Dubé, B., Lauzière, K., and Gaboury, D., 1992: Preliminary report on the structural control of the Rendell-Jackman gold deposit, Springdale Peninsula, Newfoundland; in *Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 1-10.*

Abstract

The Rendell-Jackman gold deposit is one of the most significant mesothermal vein-type gold deposits in the Canadian Appalachians. It is located within a deformed Ordovician metavolcanic dominated sequence containing felsic dykes.

The first episode of deformation is evidenced in the metavolcanics by an S_1 *planar fabric containing a down dip stretching lineation. The second episode* (D_2) *produced steeply plunging E-W folds, an* S_2 *foliation and discrete brittle-ductile* D_2 *high strain zones, affecting the metavolcanics and the felsic dykes. These brittle-ductile* D_2 *high strain zones are particularly well developed in the hinges of* F_2 *folds.*

They are also strongly controlled by the layer anisotropy induced by the felsic dykes which are competent compared to the incompetent metavolcanics. The gold bearing quartz veins are of shear vein type and hosted within and sub-parallel to the D_2 *high strain zones. The last deformation episode produced* F_3 *sub-horizontal folds and late NE and ENE brittle faults.*

Résumé

Le gisement aurifère de Rendell-Jackman est un des plus importants dépots mésothermaux de type filonien dans les Appalaches canadiennes. Le gisement est situé dans une séquence métavolcanique d'âge ordovicien, déformée et injectée de dykes felsiques.

La première phase de déformation se caractérise par le développement dans la séquence de roches métavolcaniques, d'une foliation S_1 *contenant une linéation d'étirement à fort plongement. La deuxième phase* (D_2) *a produit des plis d'orientation est-ouest à fort plongement, une foliation* S_2 *et des zones discrètes* D_2 *fragiles et ductiles très déformées, présentes dans les métavolcaniques et les dykes felsiques. Ces dernières sont particulièrement bien développées dans les charnières des plis* P_2.

Ces zones sont également fortement contrôlées par l'anisotropie de couches induite par les dykes felsiques compétents par rapport aux roches métavolcaniques incompétentes. Les veines de quartz aurifères sont de type «cisaillement» et sont contenues et orientées parallèlement aux zones D_2 *très déformées. La dernière phase de déformation a produit des plis* P_3 *sub-horizontaux et des failles cassantes tardives, d'orientation nord-est et est-nord-est.*

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[1] Published with the authorization of Noranda Exploration Co. Ltd

[2] Contribution to Canada-Newfoundland Corporation Agreement on Mineral Development 1990-1994. Project C1.124, carried out by the Quebec Geoscience Centre of the Geological Survey of Canada.

[3] Département de Géologie, Université Laval, Sainte-Foy, Québec, G1K 7P4

Structural geology of a crustal scale fault zone: the Cape Ray Fault coastal section, southwestern Newfoundland[1]

B. Dubé[2], K. Lauzière[2], and A. Tremblay[3]

Dubé, B., Lauzière, K., and Tremblay, A., 1992: Structural geology of a crustal scale fault zone: the Cape Ray Fault coastal section, southwestern Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 199-209.

Abstract

The coastal section of the Cape Ray Fault (CRF) exposes a crustal scale fault zone. Kinematic indicators in non-coaxial fabrics indicate that high grade amphibolitic deeper crustal-level rocks of the Port Aux Basques gneiss were thrust upon greenschist-grade supracrustal rocks of the volcano-sedimentary Windsor Point Group (WPG). The kinematics of the deformation is complex, with evidence of both reverse-sinistral overthrusting of the amphibolitic rocks over greenschist rocks, and reverse-dextral thrusting recorded in the WPG rocks as suggested by the variation in the stretching lineation trend and timing relationship. The structures recorded within WPG present complex internal geometry, typical of mylonitized rocks and could be attributed to a single progressive deformational event.

The discovery of a wide zone of CO_2 metasomatism within the mylonite zone may have important gold metallogenic implications since it could represent a fossil hydrothermal discharge zone across the CRF.

Résumé

La zone côtière de la faille de Cape Ray (FCR) représente une section à travers une faille crustale. Les indicateurs cinématiques développés dans les fabriques non-coaxiales indiquent que les amphibolites de niveau crustal profond, appartenant au gneiss de Port-aux-Basques, ont été chevauchés sur des roches supracrustales du faciès des schistes verts. La cinématique de la déformation est complexe comme lsemble d'indiquer la variation d'orientation des linéations d'étirement et les relations chronologiques. On constate des indices de chevauchement inverse et senestre des roches amphibolitiques sur les roches du faciès des schistes verts et, subséquemment, un chevauchement à mouvement inverse et dextre relevé à l'intérieur des roches du Groupe de Windsor Point (GWP) . Les structures enregistrées dans le GWP présentent une géométrie interne complexe, typique des zones de mylonites et peuvent être attribuées à une seule phase de déformation progressive.

La découverte au sein de la FCR d'une large zone ayant subie un métasomatisme en CO_2 peut avoir des implications importantes du point de vue de la métallogénie aurifère, étant donné qu'elle peut marquer l'emplacement la localisation d'une zone fossile de décharge hydrothermale.

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[1] Contribution to Canada-Newfoundland Cooperation Agreement or Mineral Development 1990-1994. Project C1.124, carried out by Quebec Geoscience Centre of the Geological Survey of Canada.

[2] Geological Survey of Canada/Quebec Geoscience Centre, Ste-Foy-Quebec G1V 4C7

[3] INRS-Georessources/Quebec Geoscience Centre, Quebec.

Inner shelf Quaternary sediments off northeast Newfoundland[1]

Kimberley A. Jenner and John Shaw
Atlantic Geoscience Centre, Dartmouth

Jenner, K.A. and Shaw, J. 1992: Inner shelf Quaternary sediments off northeast Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p.189-198.

Abstract

The characteristics of four northeast Newfoundland inner-shelf seabed environments – Baie Verte, La Scie, Halls Bay and Hamilton Sound - are discussed in relation to the potential of the region to host marine placer gold deposits. The shallow, relatively high-energy, coastal environments at the mouth of Baie Verte, seaward of La Scie and within Hamilton Sound appear to be the most favourable environments for concentrating particulate gold. These areas contain coarse clastics, are exposed to wave action, and have yielded samples showing anomalous gold concentrations. In contrast, the deeper, quiescent setting of Halls Bay appears to be an unfavourable site for placer development.

Résumé

Les caractéristiques de quatre milieux marins de plate-forme interne au nord-est de Terre-Neuve - Baie Verte, La Scie, la baie Halls et le détroit de Hamilton - sont traitées en fonction de la possibilité de découvrir des gisements aurifères marins. Les milieux littoraux peu profonds relativement agités à l'embouchure de la baie Verte, au large de La Scie et dans le détroit de Hamilton semblent être des milieux très favorables à la concentration de particules d'or. Ces zones contiennent des roches clastiques grossières, sont exposées à l'action des vagues et ont donné des échantillons dans lesquels les concentrations en or étaient anormales. Par contre, le milieu calme, plus profond de la baie Halls semble être un endroit non favorable à la mise en place de gisements alluvionnaires.

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[1] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development 1990-1994. Project funded by the Geological Survey of Canada.

Structural control of sill-hosted gold mineralization: the Stog'er Tight gold deposit, Baie Verte Peninsula, northwestern Newfoundland[1,2]

Donna Kirkwood[3] and Benoit Dubé
Quebec Geoscience Centre, Sainte-Foy

Kirkwood, D. and Dubé, B., 1992: Structural control of sill-hosted gold mineralization: the Stog'er Tight Gold deposit, Baie Verte Peninsula, northwestern Newfoundland; in Current Research, Part D; Geological Survey of Canada, paper 92-1D, p. 211-221.

Abstract

Structural elements within the Stog'er Tight Gold deposit have been studied in order to determine the structural history of the area, the controls on gold mineralization, and its relationship to the major faults. The Stog'er Tight deposit, a mesothermal altered wallrock gold deposit, is hosted by differentiated gabbroic sills within the Lower Ordovician Point Rousse Complex cover sequence. Three phases of deformation affected the rocks. The first phase (D_1) consisted of an intense ductile shearing which produced the regional foliation (S_1). The development of quartz-carbonate veins and associated alteration and mineralization is ascribed to a late (postductile) increment of D_1. The second event (D_2) produced south-verging asymmetric folds (F_2), ductile-brittle, high-angle faults and a fracture cleavage. Late north-northeast- to northeast-trending, broad open folds (F_3) refold the older structures. This study demonstrates the structural and lithologic influence of the hosting gabbroic sill on the mineralization.

Résumé

Les éléments structuraux présents dans le dépôt aurifère de Stog'er Tight ont été étudiés dans le but de déterminer l'histoire structurale de la région, les contrôles de la minéralisation aurifère et leur relation avec les zones de failles. Le dépôt de Stog'er Tight, un dépôt mésothermal compris dans les épontes altérées, est encaissé par des filon-couches gabbroiques différenciés compris dans une séquence de couverture datant de l'Ordovicien inférieur, à savoir le complexe de Point Rousse. Trois phases de déformation ont affecté les roches de la région. La première phase (D_1) consiste en un cisaillement ductile intense qui a produit la foliation régionale (S_1), le développement de veines de quartz et carbonates et l'altération hydrothermale associée. La minéralisation est reliée à une phase tardive (tardi-ductile) de la déformation D_1. Le deuxième événement (D_2) a produit des plis (P_2) asymmétriques à vergence vers le sud, des failles ductiles fragiles fortement inclinées et un clivage de fracture. Des failles tardives nord-nord-est à nord-est et des plis ouverts (P_3) replissent les structures plus anciennes. Cette étude démontre la forte influence structurale et lithologique du filln-couche encaissant sur la minéralisation.

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[1] Published with the authorization of Noranda Exploration Co. Ltd

[2] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development 1990-1994. Project C1.124 carried out by the Quebec Geoscience Centre.

[3] Present adress: Département de géologie, Université Laval, Sainte-Foy (Québec), G1K 7P4

Quaternary geological studies, Buchans area of central Newfoundland[1]

R.A. Klassen and P.J. Henderson

Klassen, R.A. and Henderson, P.J., 1992: Quaternary geological studies, Buchans area of central Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 11-19.

Abstract

Quaternary geological studies within the Buchans (12A/15) and Badger (12A/16) map areas are designed to establish a geological framework for mineral exploration through integration of mapping, stratigraphy, and compositional analyses of glacial sediments. In this region drift prospecting has been confounded by thick glacial deposits and complex ice flow history. In the western part of the Badger area, the dominant northeastward ice flow was preceded by southward flow. Within the Red Indian Lake basin, near Buchans, regional ice flow trending northeast–southwest was superseded by flows toward the south and the north and northwest. P-forms indicate subglacial meltwater flowing both northeast and southwest. The predominant surficial deposit is till of varying thickness and morphology. It is commonly sandy, contains evidence of internal sorting, and is overlain in places by thin outwash deposits which suggests deposition from an ablating ice sheet. Near Buchans, extensive, subtill glaciofluvial and glaciolacustrine deposits occur to 320 m elevation a.s.l. They overlie mineralized bedrock, and their incorporation within overlying till complicates the compositional relationship between bedrock and surficial sediments.

Résumé

Les études géologiques du Quaternaire, dans les régions cartographiques de Buchans (12A/15) et de Badger (12A/16), ont été conçues de façon à établir un cadre géologique pour la prospection minérale, par intégration de la cartographie, de la stratigraphie, et des analyses de la composition des sédiments glaciaires. Dans cette région, la prospection glacio-sédimentaire (drift) a été rendue très difficile par la présence d'épais sédiments glaciaires et par l'évolution complexe de l'écoulement des glaces. Dans la partie ouest de la région de Badger, l'écoulement dominant des glaces vers le nord-est a été précédé d'un écoulement vers le sud. Dans le bassin du lac Red Indian, près de Buchans, l'écoulement régional des glaces dans une direction générale nord-est-sud-ouest a fait place à des écoulements vers le sud et vers le nord et le nord-ouest. Des formes P indiquent l'écoulement d'eaux de fonte sous-glaciaires à la fois vers le nord-est et vers le sud-ouest. Les dépôts de surface prédominants sont des tills d'épaisseur et de morphologie variables. Ils sont généralement sableux, contiennent les indices d'un triage interne, et sont par endroits recouverts de minces épandages fluvio-glaciaires qui pourraient indiquer la mise en place de matériaux par suite de l'ablation d'un inlandsis. Près de Buchans, de vastes dépôts fluvioglaciaires et glaciolacustres sous-jacents au till apparaissent jusqu'à 320 m d'altitude au-dessus du niveau de la mer. Ils recouvrent un substratum minéralisé, et leur incorporation au till sus-jacent complique la relation caractérisant les compositions du substratum et les sédiments superficiels.

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[1] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-1994

Geochemistry of late Proterozoic plutonic rocks from Flemish Cap, east of the Grand Banks of Newfoundland

G. Pe-Piper[1]
Atlantic Geoscience Centre, Dartmouth

Pe-Piper, G., 1992: Geochemistry of late Proterozoic plutonic rocks from Flemish Cap, east of the Grand Banks of Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 45-48.

Abstract

Whole rock geochemical analyses of drill core samples of granodiorite from Flemish Cap show that they are typical subduction-related granitoid rocks. They closely resemble latest Proterozoic granitoid plutons of the Avalon Terrane of northern Nova Scotia and southeastern Newfoundland. A reconsideration of the existing radiometric age data is also consistent with such a correlation of the Flemish Cap granodiorite with the Avalon Terrane on land.

Résumé

Les analyses géochimiques de la roche entière, sur des échantillons de carottes de sondage prélevées dans le granodiorite du secteur de Bonnet Flamand, montrent qu'il s'agit d'une roche granitoïde typique dont la formation est liée à un épisode de subduction. Elles ressemblent fortement aux plutons granitoïdes de la toute fin du Protérozoïque, situés dans le terrane d'Avalon, dans le nord de la Nouvelle-Écosse et dans le sud-est de Terre-Neuve. Un nouvel examen des données existantes sur la datation radiométrique semble également confirmer cette corrélation du granodiorite de Flemish Cap (Bonnet Flamand) avec le terrane d'Avalon sur les terres émergées.

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[1] Department of Geology, St Mary's University, Halifax, NS, B3H 3C3
EMR Research Agreement 90-4-34

An alteration and sulphur isotope study of the Pilley's Island massive sulphides, central Newfoundland[1]

Frank Santaguida[2], Mark Hannington, and Craig Jowett[2]
Mineral Resources Division

Santaguida, F., Hannington, M., and Jowett, C., 1992: An alteration and sulphur isotope study of the Pilley's Island massive sulphides, central Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 265-274.

Abstract

Massive sulphides on Pilley's Island occur within a 200 m thick, shallow-dipping sequence of altered felsic volcanics. Mineralization in the Old Pilley's Island Mine consists of massive to semi-massive pyrite and chalcopyrite with minor sphalerite. Sulphur isotope ratios range from 2.3 to 6.5 per mil. A local exposure of transported sulphides is more polymetallic and possesses sulphur isotope ratios from 1.1 to 1.8 per mil. This material is chemically and isotopically distinct from sulphides in the Old Mine and likely represents debris from the recently discovered 3B Zone. The sulphide horizons are enclosed by a zoned alteration envelope of stockwork-associated chlorite with widespread quartz-sericite ± K-feldspar. Several stages of chlorite are recognized and mineralization is most closely associated with a distinctive, Fe-rich end-member. Application of a chlorite geothermometer indicates thermally intensifying hydrothermal activity from 240°C to 360°C. Late-stage, Mg-rich chlorite appears to correlate with an influx of cold seawater during the collapse of the hydrothermal system.

Résumé

Les sulfures massifs dans l'île Pilley reposent au sein d'une séquence légèrement inclinée de 200 m d'épaisseur composée de roches volcaniques felsiques altérées. La minéralisation dans l'ancienne mine de l'île Pilley est composée de pyrite et de chalcopyrite de nature massive à semi-massive, avec un peu de sphalérite. Les rapports isotopiques du soufre varient entre 2,3 et 6,5 par mil. Un affleurement local de sulfures transportés est plus polymétallique et présente des rapports isotopiques du soufre variant entre 1,1 et 1,8 par mil. Ce matériau se différencie chimiquement et isotopiquement des sulfures de l'ancienne mine et correspond vraisemblablement à des débris de la zone 3B récemment découverte. Les horizons sulfurés sont entourés par une enveloppe d'altération zonée de chlorite associée à un stockwerk accompagné de quantités abondantes de quartz et séricite±feldspath potassique. Plusieurs stades de chlorite ont été relevés et la minéralisation serait étroitement associée à un membre terminal ferrifère distinct. L'application d'un géothermomètre dans la chlorite révèle une activité hydrothermale à température croissante, passant de 240 °C à 360 °C. La chlorite manganésifère de stade tardif semble correspondre à un apport d'eau de mer froide durant l'effondrement du système hydrothermal.

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[1] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-94

[2] Department of Earth Sciences, University of Waterloo

A reconnaissance geotraverse through southwestern Newfoundland[1]

C.R. van Staal, J.A. Winchester[2], M. Brown[3], and J.L. Burgess[3]
Continental Geoscience Division

van Staal, C.R., Winchester, J.A., Brown, M., and Burgess, J.L., 1992: A reconnaissance geotraverse through southwestern Newfoundland; in Current Research, Part D, Geological Survey of Canada, Paper 92-1D, p. 133-143.

Abstract

Polyphase deformed gneisses and schists of the Port aux Basques Complex (PaBC) occur between the Cape Ray Fault Zone and Rose Blanche. Three distinct divisions of the complex are proposed based on lithology and separated by shear zones. These are the Grand Bay, Port aux Basques and Otter Bay divisions. Rocks of the complex were affected by four generations of folds and a complex history of shearing, localized in narrow fault zones. D2 structures and syntectonic high temperature and pressure metamorphism were probably produced during orogenic thickening as a result of Silurian collision. The D3 and D4 structures are tentatively attributed to respectively sinistral and dextral transpression, which was manifested by sinistral and dextral strike-slip and (or) reverse movement along the major fault zones.

Résumé

Les gneiss et les schistes déformés au cours de plusieurs phases du complexe de Port aux Basques (CPaB) sont situés entre la zone de failles de Cape Ray et Rose-Blanche. Trois divisions distinctes du complexe sont proposées, basées sur la lithologie et séparées par des zones de cisaillement. Il s'agit des divisions de Grand Bay, de Port aux Basques et d'Otter Bay. Les roches du complexe ont été affectées par quatre générations de plis et des cisaillements complexes dans des zones faillées étroites. Les structures D_2 et le métamorphisme syntectonique à température et pression élevées ont probablement été produits durant l'épaississement orogénique causé par la collision silurienne. Les structures D_3 et D_4 ont été temporairement attribuées à une transpression respectivement senestre et dextre, qui s'est manifestée par un coulissage senestre et dextre ou un déplacement inverse le long des principales zones de failles, ou les deux.

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[1] The work was funded under Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-94

[2] Department of Geology, University of Keele, Keele, Staffordshire, ST5 5BG, England

[3] Department of Geology, University of Maryland at College Park, MD 20742, USA

Mélanges and coticule occurrences in the northeast Exploits Subzone, Newfoundland[1]

Harold Williams[2]
Continental Geoscience Division

Williams, H., 1992: M-82-langes and coticule occurrences in the northeast Exploits Subzone, Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 121-127.

Abstract

The 1991 field season was spent examining the Dog Bay Melange and coticules (spessartine garnet-quartz layers) that are common associates of melanges in the northeast Exploits Subzone.

The Dog Bay Melange is defined by large altered gabbro, mafic volcanic, and smaller sandstone blocks in an otherwise bedded section of shale and siltstone at Dog Bay Point.

Dark grey shales with thin coticule layers and nodules are associated with the Dog Bay, Carmanville, and Dunnage melanges, and they occur in an area of regionally metamorphosed rocks that extends from Aspen Cove to Ragged Harbour in the Carmanville area. Regional relationships indicate a Middle Ordovician or earlier age for the coticules and associated melanges. The intermediate position of the Dog Bay Melange between the Dunnage and Carmanville melanges, and coticule occurrences in all examples imply continuity of a coastal lithic belt for 50 km across the northeast Exploits Subzone.

Résumé

Les travaux de terrain de la saison 1991 ont été consacrés à l'étude du mélange et des coticules de Dog Bay (couches de spessartite et quartz) qui sont couramment associés aux mélanges contenus dans le nord-est de la sous-zone d'Exploits.

Le mélange de Dog Bay est composé de grands blocs altérés de gabbro, de roches volcaniques mafiques et de blocs de grès de plus petite taille dans une coupe de shale et de siltstone stratifiée à la pointe Dog Bay.

Les shales gris foncé à minces couches et nodules de coticules sont associés aux mélanges de Dog Bay, de Carmanville et de Dunnage, et ils reposent dans une zone de roches métamorphisées régionalement qui s'étend d'Aspen Cove à Ragged Harbour dans le région de Carmanville. Les liens régionaux indiquent que les coticules et les mélanges associés remontent à l'Ordovicien moyen ou même à une époque antérieure. La position intermédiaire du mélange de Dog Bay entre les mélanges de Dunnage et de Carmanville, et la présence de coticules dans tous les exemples laissent supposer une continuitédans une zone lithique littorale sur une distance de 50 km à travers le nord-est de la sous-zone d'Exploits.

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[1] Lithoprobe Contribution no. 264

[2] Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, A1B 3X5

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GEOLOGICAL SURVEY BRANCH

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